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CONTENTS

	Pages
Geology of the Lima region, southwestern Montana and adjacent Idaho. By Robert Scholten, K. A. Keenmon, and W. O. Kupsch	345–404
Distribution of igneous rocks in space and time. By G. W. Tyrrell	405–426
Provenience of pyroclastic materials. By Clarence S. Ross	427–434
Geology of Catoctin Mountain, Maryland and Virginia. By John C. Whitaker	435–462
Short Notes	
Pillar Falls mud flow and Shoshone Falls andesite near Twin Falls, Idaho. By Harold T. Stearns	463–464
Toro Seamount. By J. Lamar Worzel and G. Lynn Shurbet	465–466
Discovery of *Sonoraspis* in Southern California. By Alexander Stoyanow and Takeo Susuki	467–470

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PAPERS IN PRESS FOR FORTHCOMING ISSUES

ORIGIN OF THE UPLAND SILT NEAR FAIRBANKS, ALASKA. By Troy L. Péwé

PALYNOLOGY IN SOUTHERN NORTH AMERICA

INTRODUCTION AND ACKNOWLEDGMENTS. By Paul B. Sears

PART II. STUDY OF TWO CORES FROM LAKE SEDIMENTS OF MEXICO CITY BASIN. By Fred Foreman

PART III. MICROFOSSIL PROFILES UNDER MEXICO CITY CORRELATED WITH THE SEDIMENTARY PROFILE. By Kathryn H. Clisby and Paul B. Sears

PART IV. PLEISTOCENE CLIMATE IN MEXICO. By Paul B. Sears and Kathryn H. Clisby

GEOMORPHOLOGY OF SOUTH-CENTRAL WASHINGTON, ILLUSTRATED BY THE YAKIMA EAST QUADRANGLE. By Aaron C. Waters

GEOLOGY OF CATHEDRAL MOUNTAIN QUADRANGLE, BREWSTER COUNTY, TEXAS. By William N. McAnulty

GEOLOGY OF THE BLUE RIDGE REGION IN CENTRAL VIRGINIA. By Robert O. Bloomer and Harry J. Werner

TRANSFORMATION OF INDICES IN TWIN GLIDING. By A. Pabst

PLEISTOCENE GEOLOGY OF THE SOUTHWESTERN WIND RIVER MOUNTAINS, WYOMING. By G. William Holmes and John H. Moss

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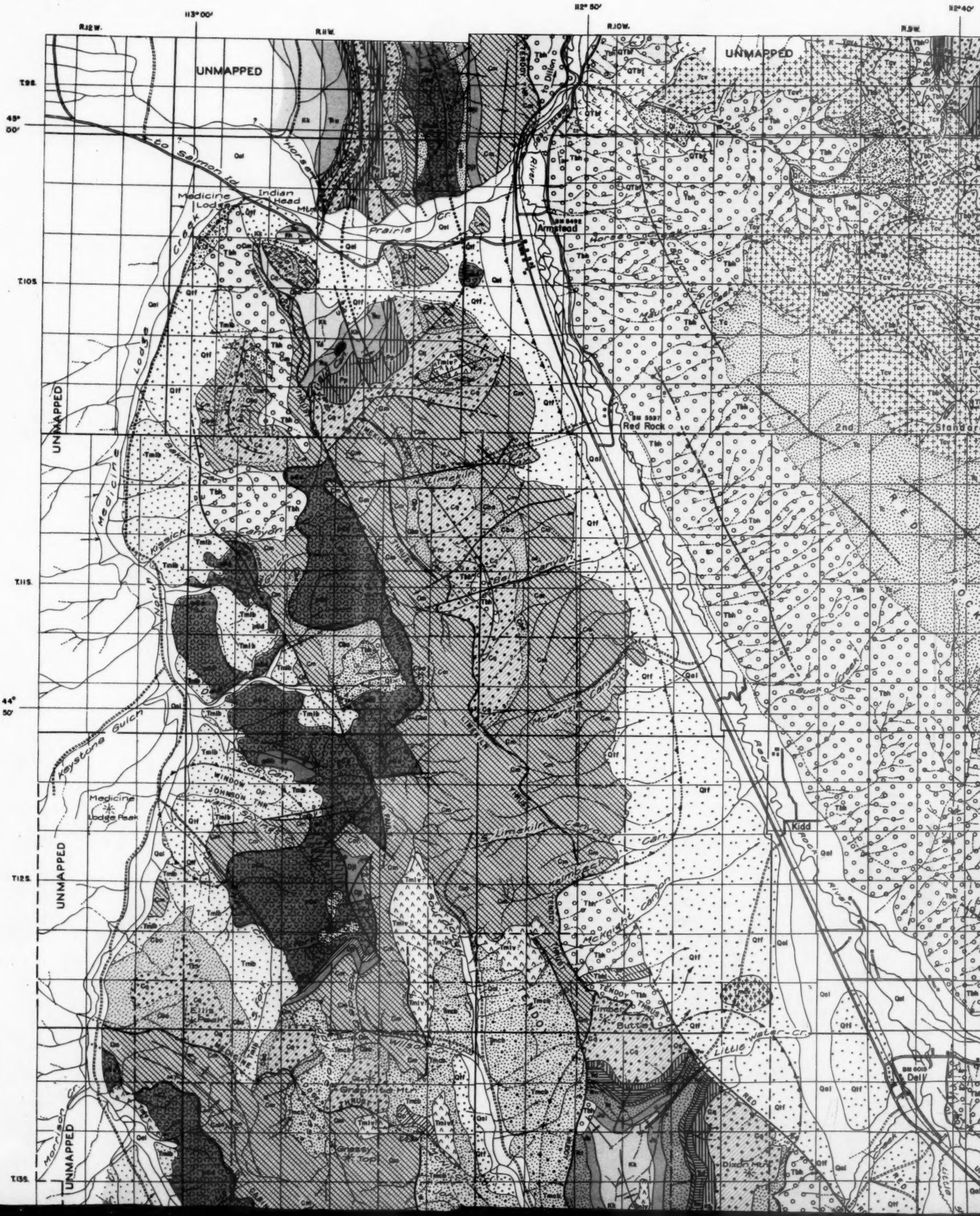
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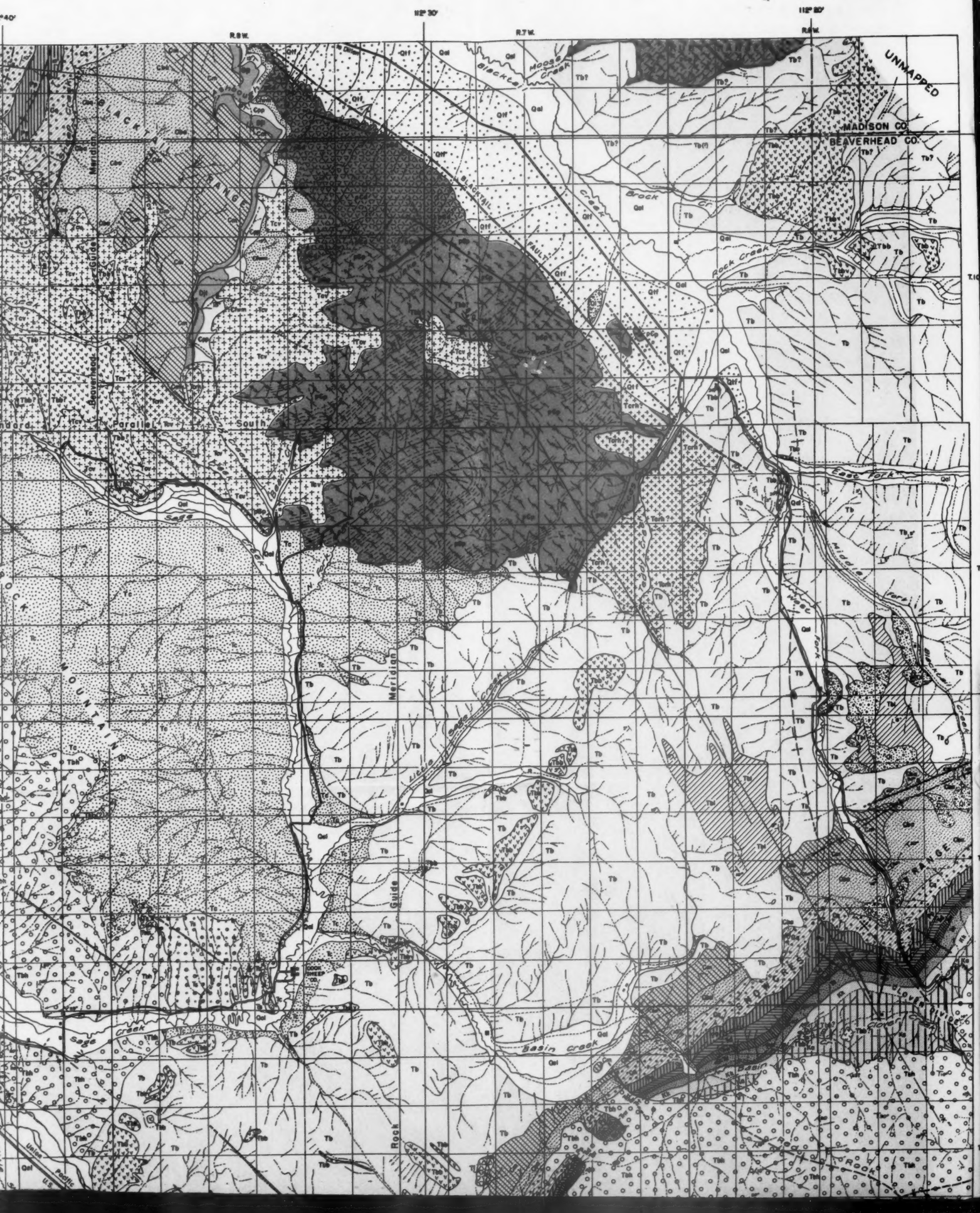
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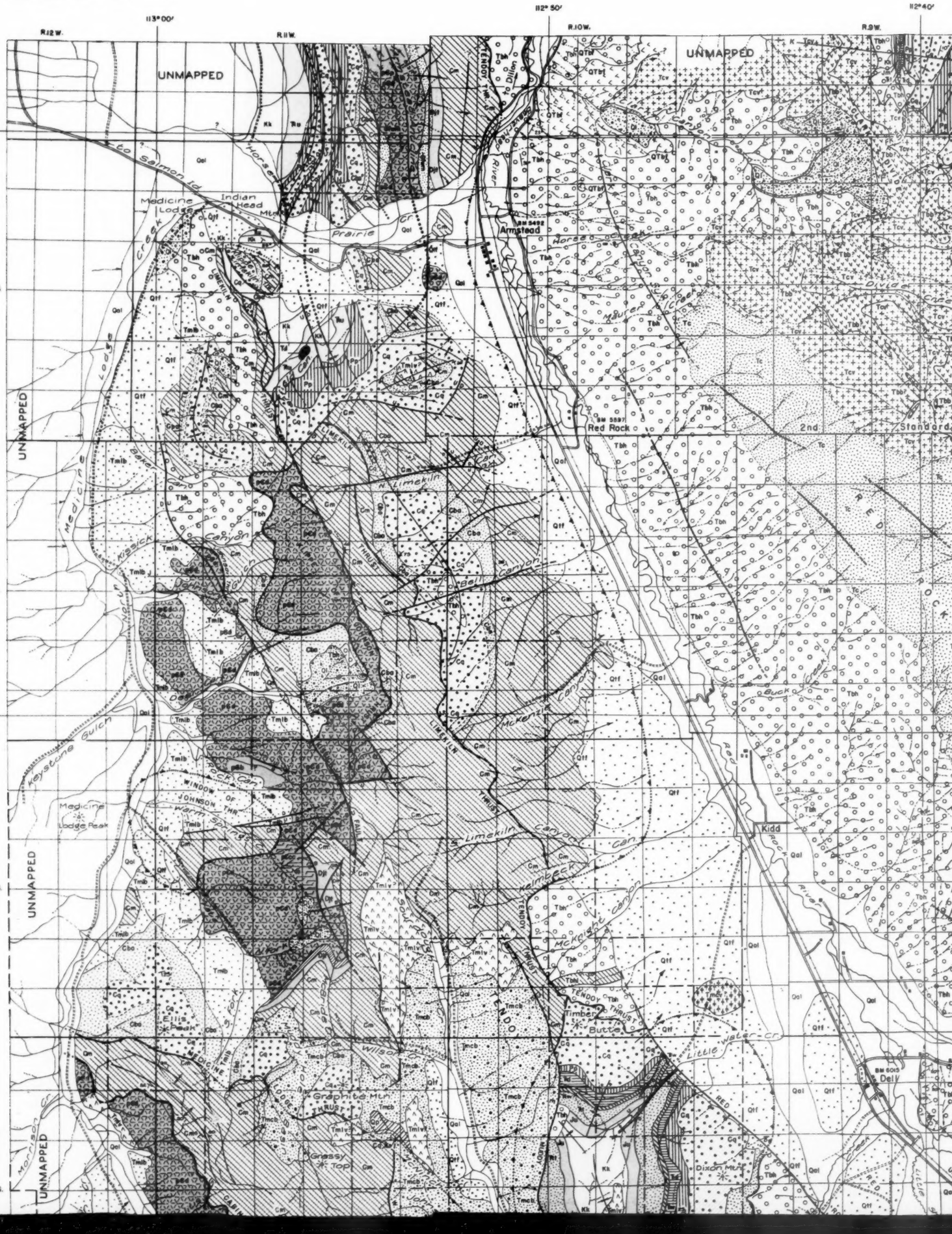
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Warm Spring Creek
Kidd
Timber Butte
Little Water Cr.
Graphite Mtn.
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Ellis Peak
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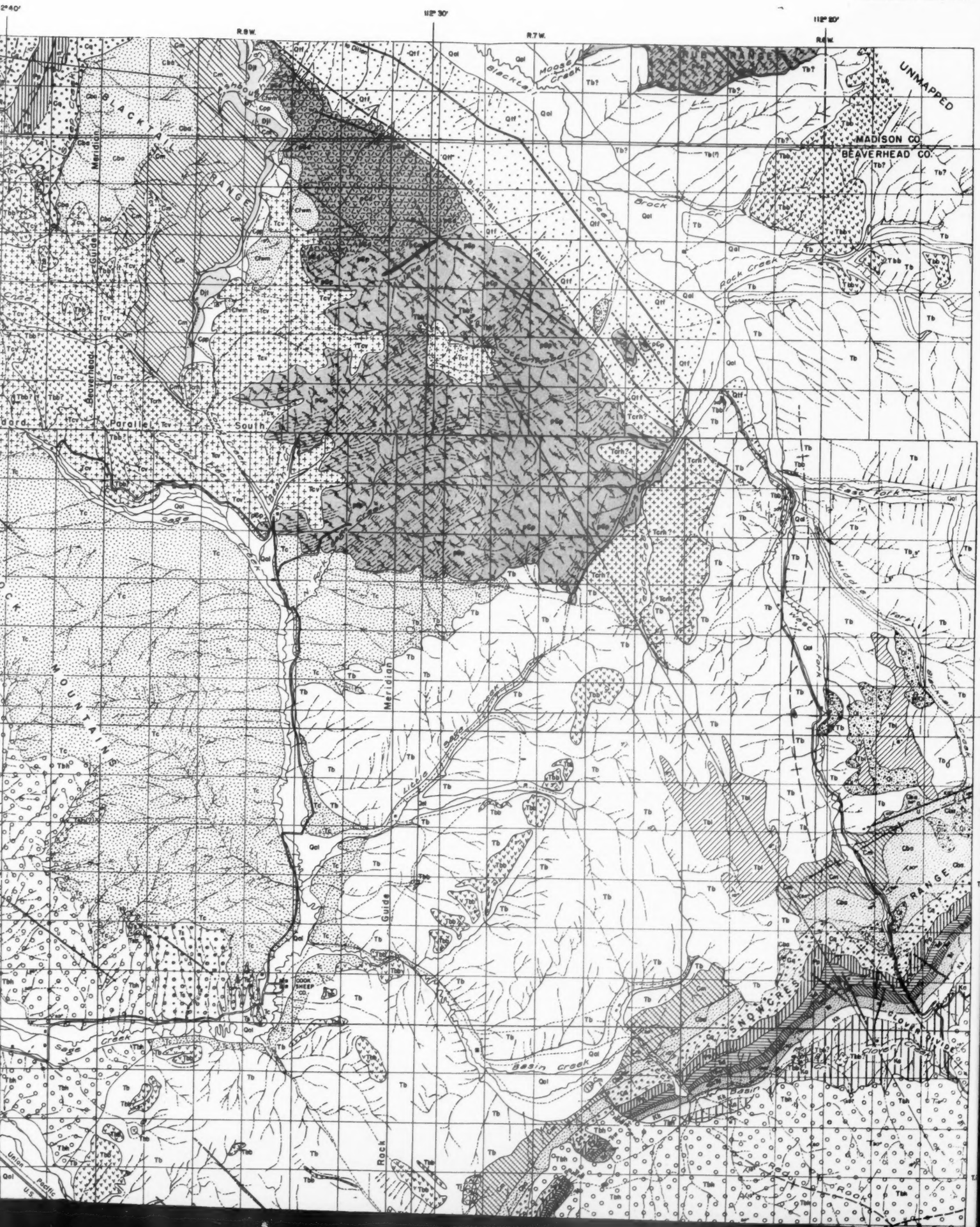
Tendoy Thrust
Johnson Thrust
Limekiln Thrust
Window of Johnson Thr.
Red Rock River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Blacktail Deer Creek Fault
Moose Creek
Brock Cr.
Rock Creek
Little Sage Creek
Middle Fork
East Fork
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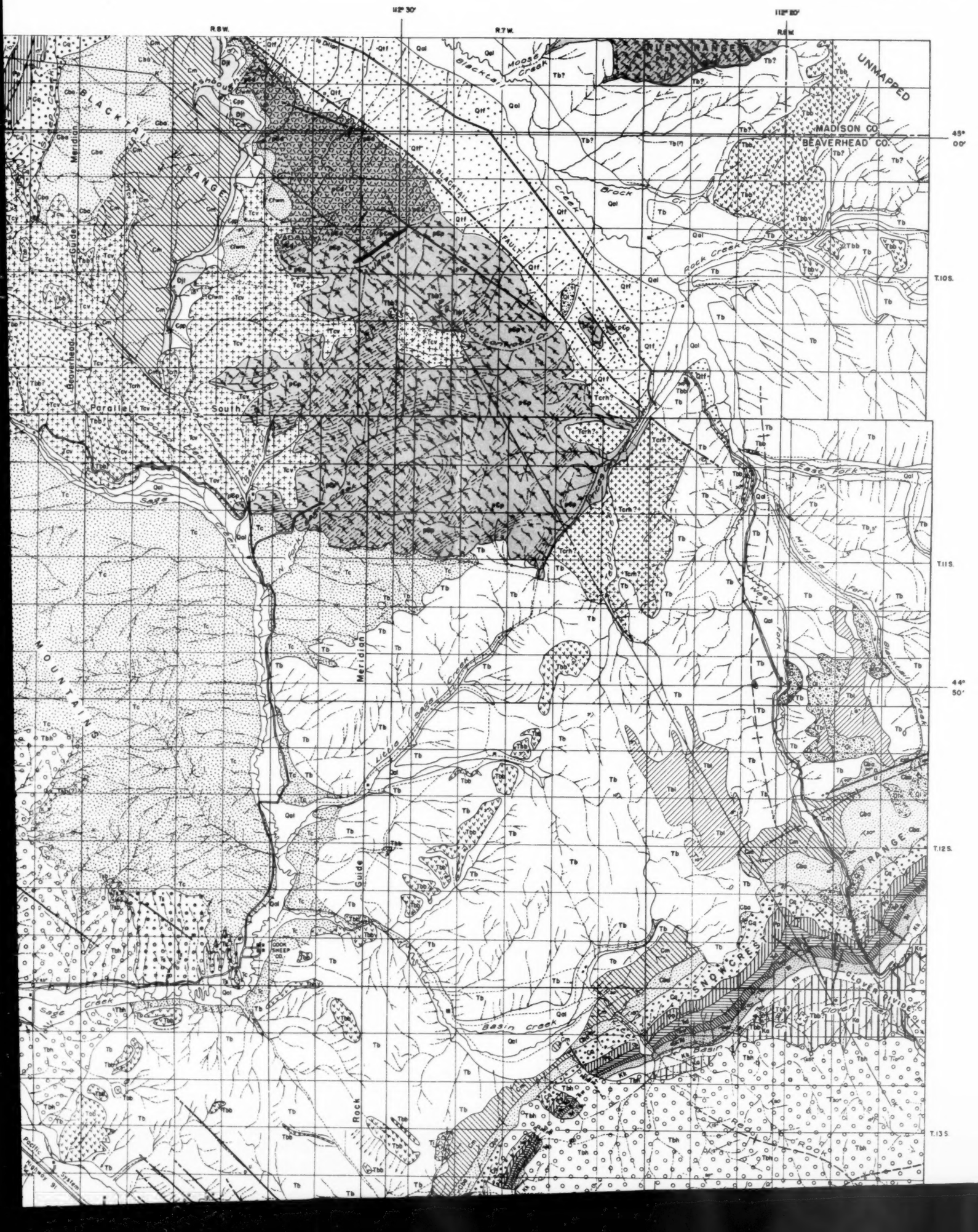
Beaverhead Co.
Unmapped
Cook Sheep Co.
Union Pacific
U.S.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Beaverhead Meridian
Guide
Parallel
South
MOUNTAIN
Meridian
Rock
Pacific Highway 91
Moose Creek
Blacktail Creek
Rock Creek
Cottonwood Creek
Sage Creek
Little Sage Creek
East Fork
Middle Fork
Basin Creek
Clover Creek
SNOWCREST RANGE
GOOK SHEEP CO.

EXPLANATION

SEDIMENTARY ROCKS

QUATERNARY

Recent

Qal — Alluvium

Ql — Landslides

Qtf — Terrace gravels and alluvial fans

UNCONFORMITY

Pleistocene

Qp — "Pinedale"? glacial deposits

UNCONFORMITY

Qb — "Buli Lake"? glacial deposits

UNCONFORMITY

TERTIARY

Miocene

West and south of U.S. 91: Tbru, Tmlb — "Medicine Lodge" beds (Tbru: includes undifferentiated rhyolites)

East and north of U.S. 91: Tbl, Tb — Blacktail Deer Creek formation (Tbl: fresh water limestone)

UNCONFORMITY

Oligocene

Tmcb — "Muddy Creek" beds

Tc — Cook Ranch formation

UNCONFORMITY

Eocene

Ts — Sage Creek formation

UNCONFORMITY

Paleocene

Tbh, Tbhl — Beaverhead formation (Tbhl: fresh water limestone)

UNCONFORMITY

CRETACEOUS

Ka — Aspen formation

Kk — Kootenai formation

UNCONFORMITY

JURASSIC

Jm — Morrison formation

Jr — Rierdon formation

Ju — Jurassic undifferentiated

Js

PRE-CAMBRIAN

p€b — Belt group

METAMORPHIC ROCKS

UNCONFORMITY

p€d — Dillon granite gneiss

p€cc — Cherry Creek group

p€p — Pony group

IGNEOUS ROCKS

EXTRUSIVES

QUATERNARY

Pleistocene to Recent

West and south of U.S. 91: QTbl — Snake River basaltic lava flows

East and north of U.S. 91: QTbf — Clark Canyon basaltic lava flows

TERTIARY

Pliocene

Tesr — Edie School rhyolitic lava flows

Miocene

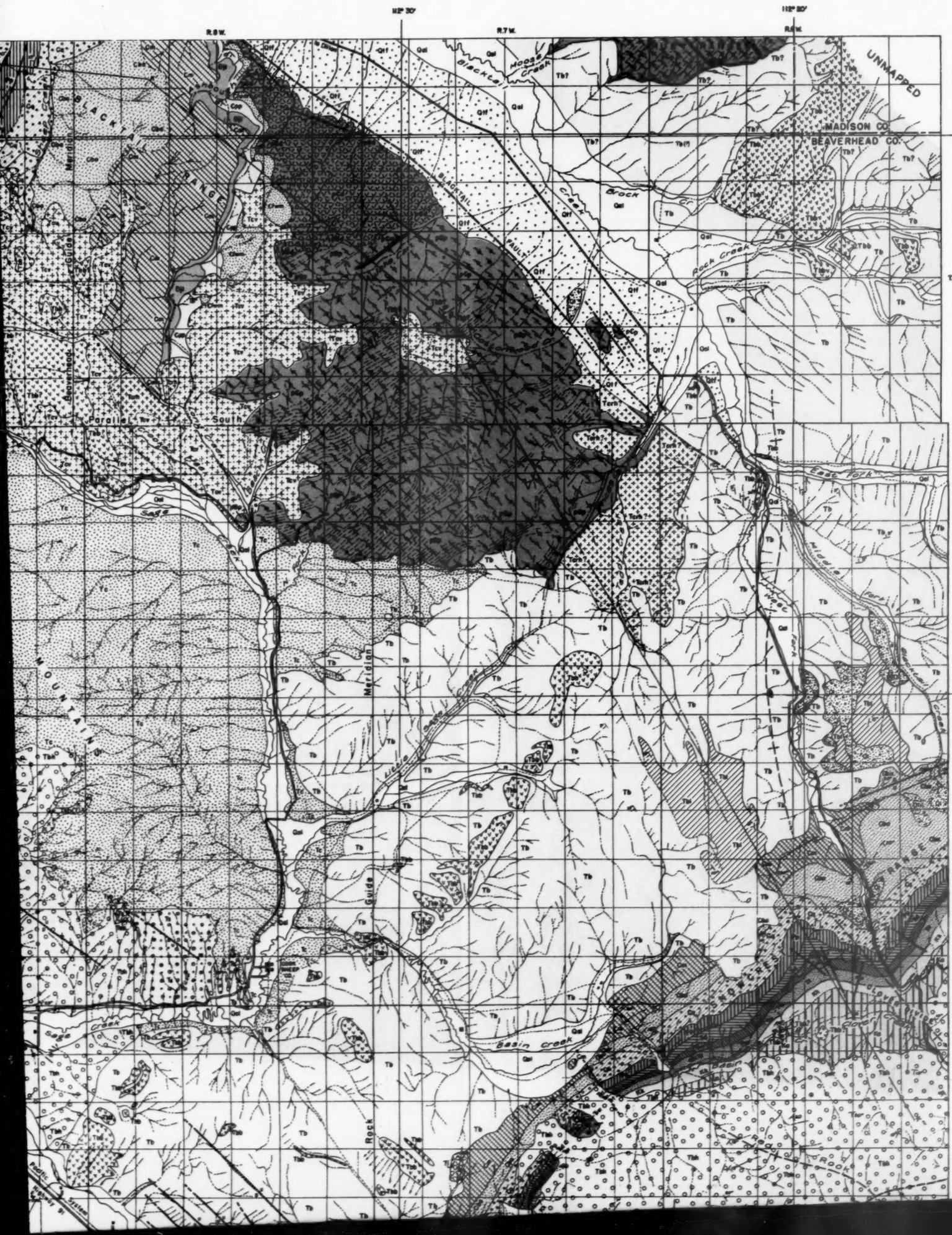
Tbb — Blacktail Deer Creek basaltic lava flows (Tbb?: probably correlative basaltic and locally andesitic lava flows)

Tmlv — Medicine Lodge volcanics (basaltic and andesitic lavas with minor tuffs and agglomerates)

Oligocene

Tcrh — Cook Ranch rhyolitic lava flows (Tcrh?: probably correlative lavas and pyroclastics)

Tmcv — Muddy Creek volcanics (rhyolitic tuffs and lava flows)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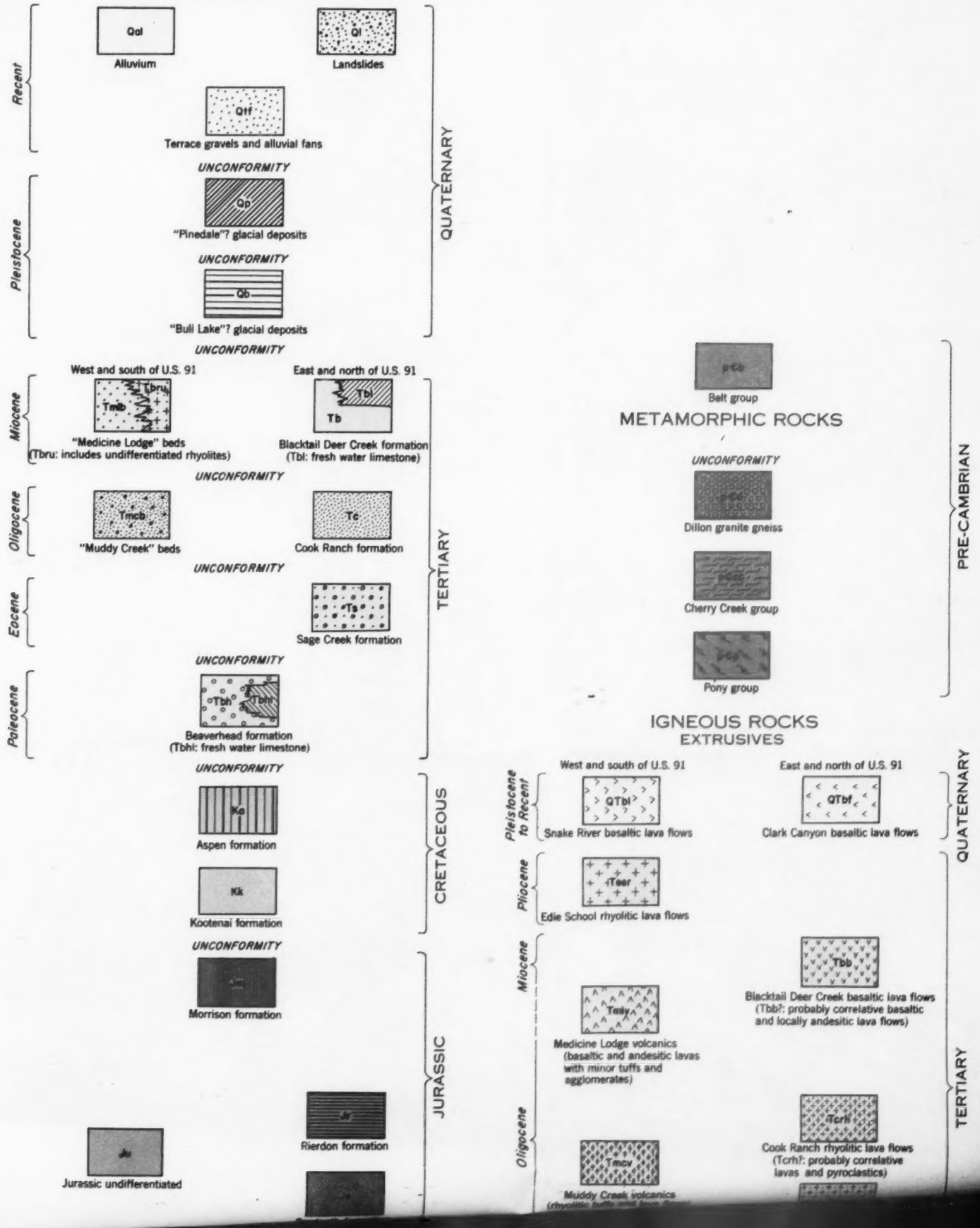
BLACKTAIL RANGE
MOUNTAIN
Blacktail
Hoose Creek
Brock Creek
Rock Creek
BLACKTAIL FAULT
Parallel
South
Sage
Creek
Meridian
Little Sage Creek
Guide
Rock
Basin Creek
East Fork
Middle Fork
Pacific
Power & Light
System

EXPLANATION

SEDIMENTARY ROCKS



METAMORPHIC ROCKS

IGNEOUS ROCKS

EXTRUSIVES

MAP SHOWING LOCATION OF AREA

I, KEENMON; II, KUPSCH; III SCHOLTEN

Township and section base taken from U.S. Forest Service Map of Beaverhead National Forest, 1947. Geology mapped on aerial photographs of the U.S. Department of Agriculture, Agricultural Adjustment Agency, 1942. Section lines were established on the aerial photographs by section corners and various land lines and then transfered to the Forest Service base map. Drafting by D.L. Bell.

GEOLOGIC MAP OF THE LI

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LIMA REGION, MONTANA-IDAHO

SCALE

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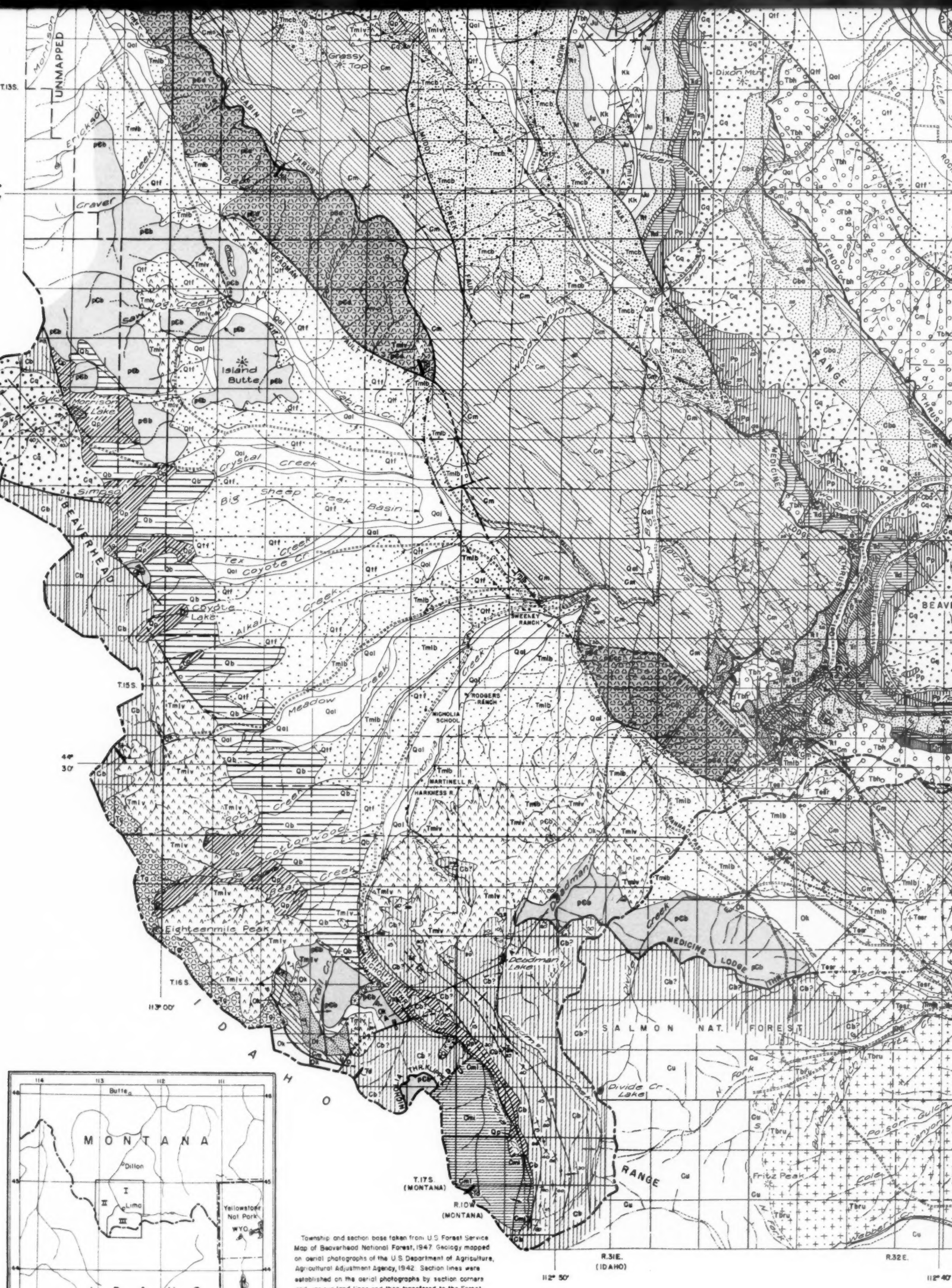

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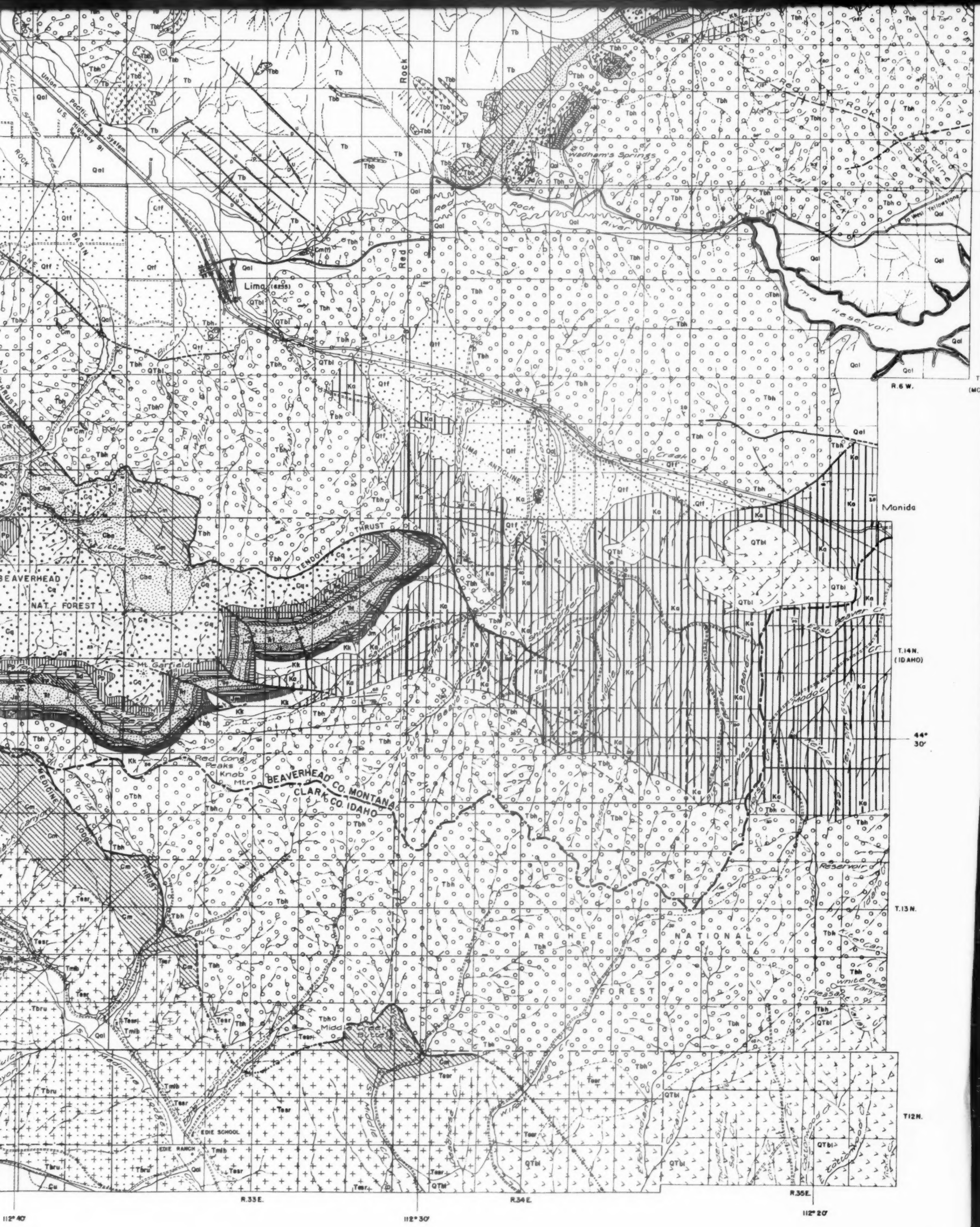

LIMA REGION, MONTANA-IDAHO

SCALE

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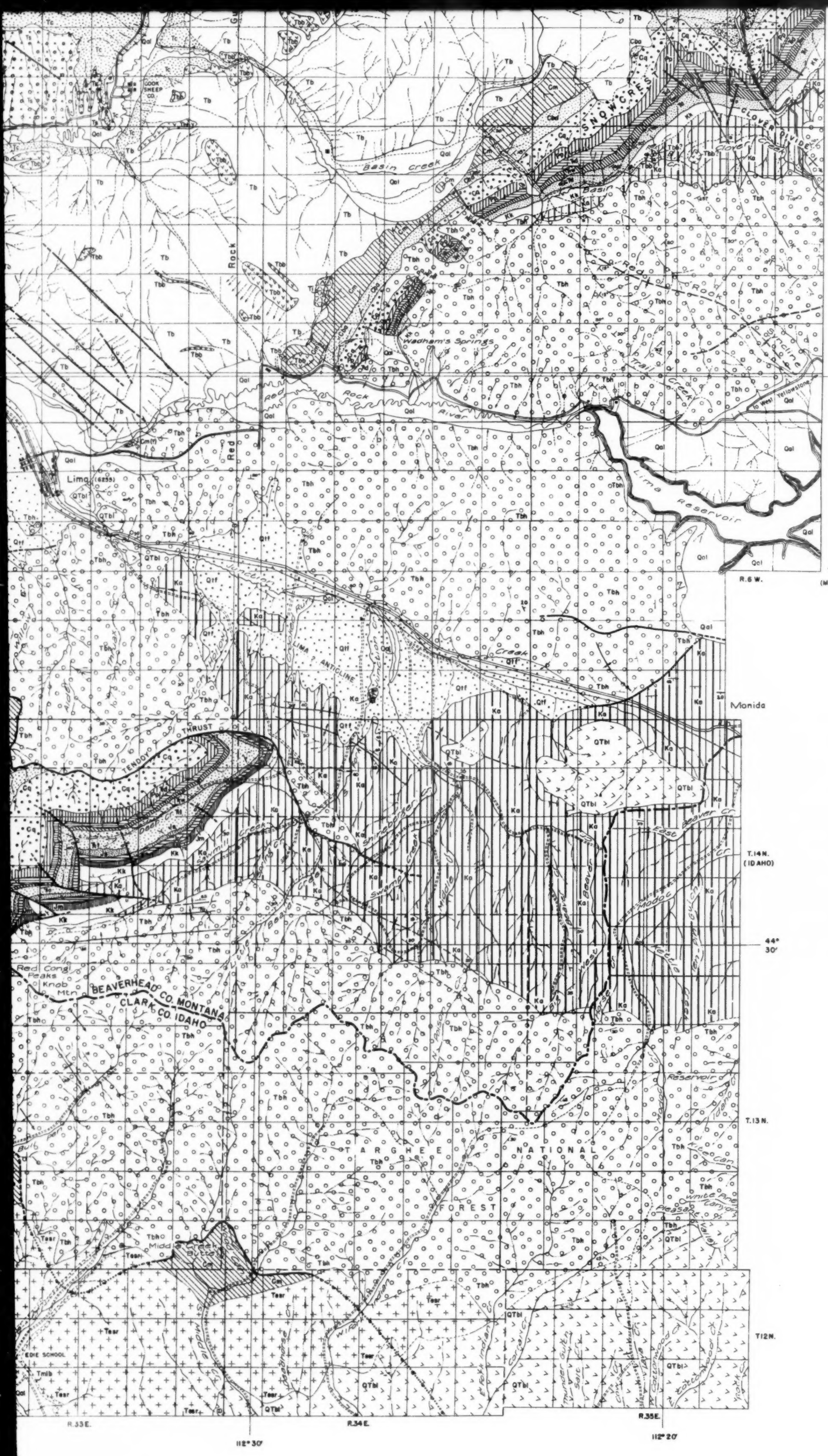


Jurassic un

Triassic un

Pennsylvanian

Mississippian

Mississippian

Jefferson and Thr

Flathead and Mea

ONTANA-IDAHO

Kk
Kootenai formation

UNCONFORMITY

Jm
Morrison formation

Ju
ssic undifferentiated

Jr
Rierdon formation

Js
Sawtooth formation

UNCONFORMITY

Tu
ssic undifferentiated

Tt
Thaynes formation

Tw
Woodside formation

Td
Dinwoody formation

Pp
Phosphoria formation

Cq
Quadrant formation

Cbs
Big Snowy and Amsden formations

Cu
pian undifferentiated

Cb
Brazer formation

Cmi
Milligen formation

Cm
Madison group - Mission Canyon and Lodgepole formations

DJt
nd Three Forks formations

Dt
Three Forks formation

UNCONFORMITY

Ok
Kinnikinic formation

UNCONFORMITY

€pp
Park and Pilgrim formations

€fm
d Meagher formations

€fwm
Flathead, Wolsey and Meagher formations

UNCONFORMITY

CRETACEOUS

JURASSIC

TRIASSIC

PERMIAN

CARBONIFEROUS

DEVONIAN

ORDOVICIAN

CAMBRIAN

Pliocene

Tesr
Edie School rhyolitic lava flows

Miocene

Tmlv
Medicine Lodge volcanics
(basaltic and andesitic lavas with minor tuffs and agglomerates)

Tbb
Blacktail Deer Creek basaltic lava flows
(Tbb?: probably correlative basaltic and locally andesitic lava flows)

Oligocene

Tmcv
Muddy Creek volcanics
(rhyolitic tuffs and lava flows)

Tcrh
Cook Ranch rhyolitic lava flows
(Tcrh?: probably correlative lavas and pyroclastics)

Tcv
Cook Ranch volcanics
(rhyolitic tuffs and agglomerates)

Eocene

Tsb
Sage Creek basaltic lava flows
(Tsb?: probably correlative basaltic lavas)

TERTIARY

INTRUSIVES

Eocene

Tg
Beaverhead granite

Undifferentiated

Ti
Granitic intrusive

Td
Dikes

TERTIARY

p€dd
Diabase dike

p-€m
Ultra-mafic sheet

PRE-CAMBRIAN

SYMBOLS

FOLDS

Anticline

Syncline

FOLIATION

(shown in Blacktail Range only)

SURFACE OPENINGS

Trench

Pit

FAULTS

U

D

Normal

Thrust

Inferred thrust
(sawteeth on side of upper plate)

Concealed or indefinite thrust

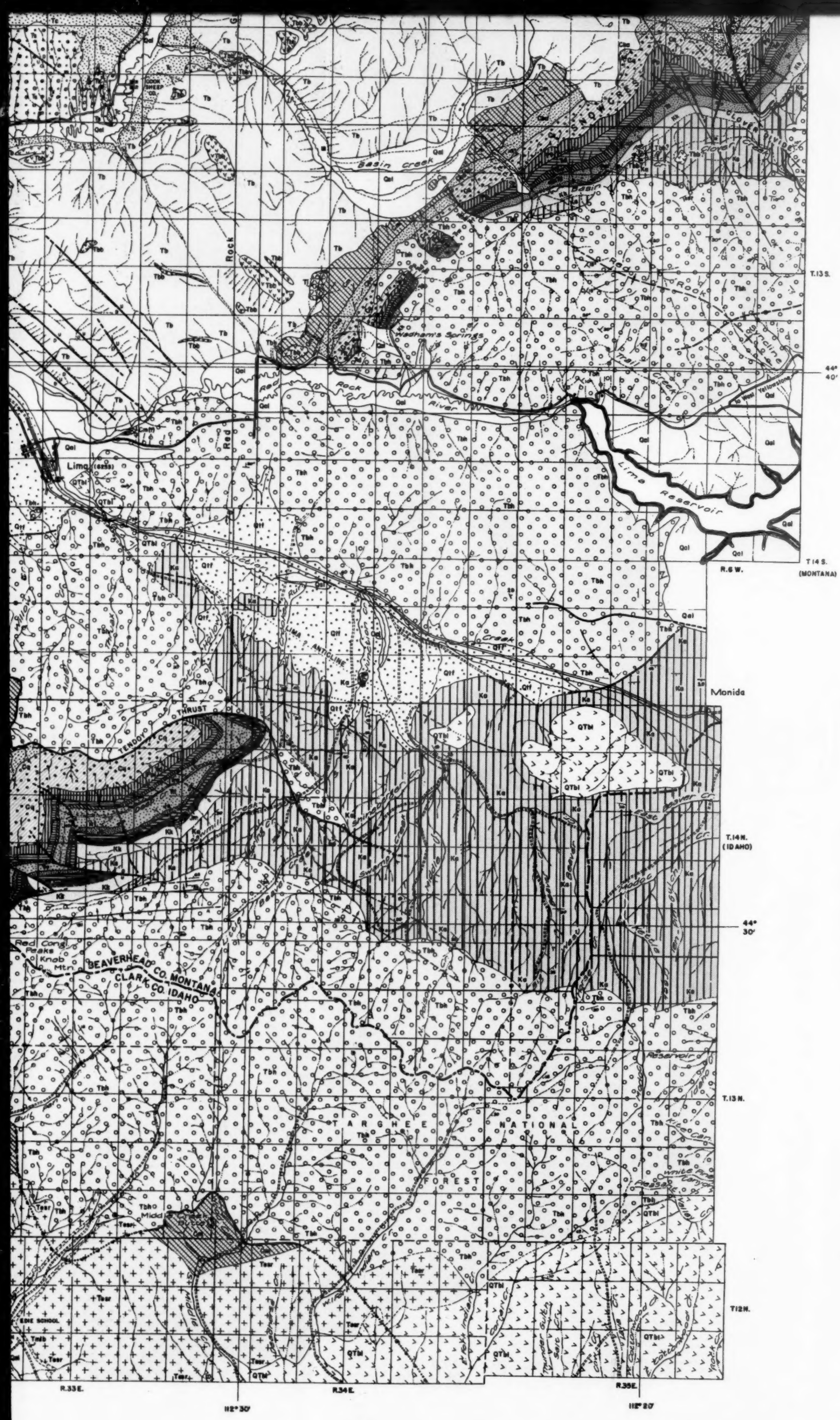
Field work: July 15-August 22, 1947, and June 14-September 6, 1948

19½°

true north

magn. north

Approximate mean declination; 1935



ONTANA-IDAHO

Kk
Kootenai formation

UNCONFORMITY

Jm
Morrison formation

Jr
Rierdon formation

Ju
Jurassic undifferentiated

Js
Sawtooth formation

UNCONFORMITY

Tt
Thaynes formation

Tw
Woodside formation

Tru
Triassic undifferentiated

Td
Dinwoody formation

Pp
Phosphoria formation

Cq
Quadrant formation

Cbs
Big Snowy and Amsden formations

Cb
Brazer formation

Cu
Cmi
Mississippian undifferentiated

Milligen formation

Cm
Madison group - Mission Canyon and Lodgepole formations

Dgt
Dt
n and Three Forks formations

Three Forks formation

UNCONFORMITY

Ok
Kinnikinic formation

UNCONFORMITY

Cpp
Park and Pilgrim formations

Cfm
Cfwm
d and Meagher formations

Flathead, Wolsey and Meagher formations

UNCONFORMITY

CRETACEOUS

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Tear
Edie School rhyolitic lava flows

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Tcrh
Cook Ranch rhyolitic lava flows
(Tcrh?: probably correlative lavas and pyroclastics)

Tcv
Cook Ranch volcanics
(rhyolitic tuffs and agglomerates)

Eocene

Tsb
Sage Creek basaltic lava flows
(Tsb?: probably correlative basaltic lavas)

TERTIARY

INTRUSIVES

Eocene

Tg
Beaverhead granite

Undifferentiated

Tli
Granitic intrusive

Td
Dikes

TERTIARY

pCdd
Diabase dike

pCm
Ultra-mafic sheet

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Concealed or indefinite thrust

(sawteeth on side of upper plate)

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19½°

true north

magn. north

Approximate mean declination; 1935

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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 66, PP. 345-404, 6 FIGS., 6 PLS. APRIL 1955

GEOLOGY OF THE LIMA REGION, SOUTHWESTERN MONTANA AND ADJACENT IDAHO

BY ROBERT SCHOLTEN, K. A. KEENMON, AND W. O. KUPSCH

ABSTRACT

The Lima region which includes the southwest corner of Beaverhead County, Montana, and an adjacent portion of Clark County, Idaho, covers most of the Tendoy, Blacktail, and Snowcrest ranges, parts of the Beaverhead and Ruby ranges, and the intermontane basins.

Metamorphic rocks include the Pony and Cherry Creek groups and the Dillon granite gneiss, all pre-Beltian. The sedimentary rocks represent all systems from Beltian through Quaternary, except the Silurian. The pre-Cretaceous rocks were deposited in a tectonic shelf environment and in adjacent parts of the miogeosyncline to the west. This tectonic differentiation was especially marked during the Mississippian. The latest Jurassic and post-Jurassic sediments are terrestrial deposits related to the Nevadan and Laramide orogenies. A Laramide granite stock crops out at the western margin of the region. Extrusive rocks include basaltic and rhyolitic lavas and pyroclastics, ranging in age from Eocene through Pleistocene.

Pre-Beltian diastrophism metamorphosed the Pony and Cherry Creek rocks and the partly syntectonic Dillon granite complex. The Laramide orogeny began with northeast folding, probably Paleocene, which resulted in deposition of the thick and coarse Beaverhead conglomerate. This was followed in early Eocene time by northwest folding and overthrusting to the northeast. The low-angle Medicine Lodge thrust, which may have had a displacement of 10 miles or more, carried the geosynclinal sedimentary facies close to the shelf facies. The Johnson and Limekiln thrusts probably are branches of this fault. These low-angle thrusts were subsequently broken, dislocated, and tilted by the high-angle Nicholia, Cabin, and Tendoy thrusts. Post-Laramide diastrophism consisted mostly of block faulting, which may have started in mid-Tertiary time and, in part, continues. It disrupted a late Tertiary peneplain and in part formed the present ranges. Northeast drainage on this surface maintained itself across the Tendoy Range and cut several deep gorges. Fresh-water sediments and volcanic material accumulated in the basins and were tilted.

Glaciers of the Wisconsin stage shaped some of the valleys, depositing drift in two substages. The present cycle of erosion is in the mature stage.

Mineral resources in the region are largely undeveloped. One small lead-zinc mine is in operation, but in general possibilities for metal mining are poor. Coal and gypsum are, or have been, mined, and phosphate and associated elements may become important in the future. The region is not believed to have major oil and gas potentialities.

CONTENTS

	Page
Introduction	346
Location	346
Previous work	346
Present study	348
Acknowledgments	348
Geography	348
Topographic features	348
Drainage	348
Climate and vegetation	349
Accessibility	350
Metamorphic rocks	350
Introduction	350
Pre-Beltian System	350
Pony group	350
Cherry Creek group	350
Dillon granite gneiss	351
Sedimentary rocks	352
Introduction	352
Stratigraphic column	352
Paleotectonic control	352
Paleogeographic control	353
Beltian System	353
Cambrian System	353
Ordovician System	363

	Page
Devonian System	363
Mississippian System	363
Pennsylvanian System	366
Permian System	366
Triassic System	366
Jurassic System	367
Cretaceous System	367
Tertiary System	368
Quaternary System	370
Igneous rocks	370
Introduction	370
Intrusive rocks	370
Beaverhead granite	370
Snowcrest granite	372
Hypabyssal bodies	372
Extrusive rocks	373
Sage Creek basalts	373
Cook Ranch rhyolites	373
Cook Ranch volcanics	374
Blacktail Deer Creek basalts	374
Muddy Creek volcanics	374
Medicine Lodge volcanics	375
Edie School rhyolites	376
Snake River basalts	377
Clark Canyon lavas	377
Structure	377
Pre-Beltian structures	377
Laramide structures	378
General remarks	378
Northeast folds	378
Northwest folds	380
Other folds	381
Low-angle thrusts	382
High-angle thrusts	383
Other Laramide faults	384
Post-Laramide structures	385
General remarks	385
Folds	385
Block faults	385
Other post-Laramide faults	387
Regional relations	389
Geomorphology	389
Mountain ranges	389
Stream valleys	389
Erosion surfaces	391
General remarks	391
Old surface	391
Intermediate surface	391
Young surface	392
Glacial features	392
Erosional features	392
Depositional features	393
Regional relations	393
Geologic history	394
Precambrian events	394
Paleozoic events	394
Mesozoic events	395
Tertiary events	396
Quaternary events	398
Economic geology	399
Metallic resources	399
Nonmetallic resources	399
Phosphate	399
Gypsum	400
Lime	400
Bentonite	400
Gravel	400
Building stone	400
Graphite	400
Coal	400
Petroleum and natural gas	400
Ground water	401
References cited	401

ILLUSTRATIONS

Figure	Page
1. Location map	347
2. Geomorphic diagram	349
3. Mid-Laramide tectonic map	379
4. Late Laramide tectonic map	382
5. Post-Laramide tectonic map	386
6. Composite map of Cenozoic tectonic elements	388

Plate	
1. Geologic map of the Lima region	Facing page 345
2. Features of sedimentary rocks	Following page 364
3. Paleozoic carbonate rocks	
4. Structural features	
5. Geomorphic features	
6. Geologic cross sections of the Lima region	Facing page 403

TABLES

Table	Page
1. Sedimentary rocks in the Lima region	354
2. Permo-Carboniferous correlation chart	364
3. Igneous rocks in the Lima region	371

INTRODUCTION

Location

The Lima region is in the southern part of Beaverhead County, southwestern Montana, and extends into Clark County, east-central Idaho (Fig. 1). The area comprises about 1600 square miles.

Previous Work

The first scientific explorers to pass through the area were Lewis and Clark, who in August 1805 camped close to the present site of Armstead. The explorers went up Horse Prairie Creek which they erroneously believed to be the ultimate source of the Missouri.

The first geological observations on the Lima region were published in 1872 as a part of the fifth annual report of the Hayden Survey (Hayden, 1872, p. 144–146). This party noted the anticlinal valley north of Horse Prairie Creek and assigned its core to the Precambrian. Another early investigator is Douglass (1902; 1903), who collected and described vertebrate fossils from Tertiary beds in the

eastern part of the region. Bowen (1917) studied the phosphatic oil shales in the Permian Phosphoria formation. In a geological reconnaissance of central Idaho, Kirkham (1927, p. 21–26) mentions the coarse, red conglomerate and the Medicine Lodge overthrust exposed in the Idaho portion of the Lima region.

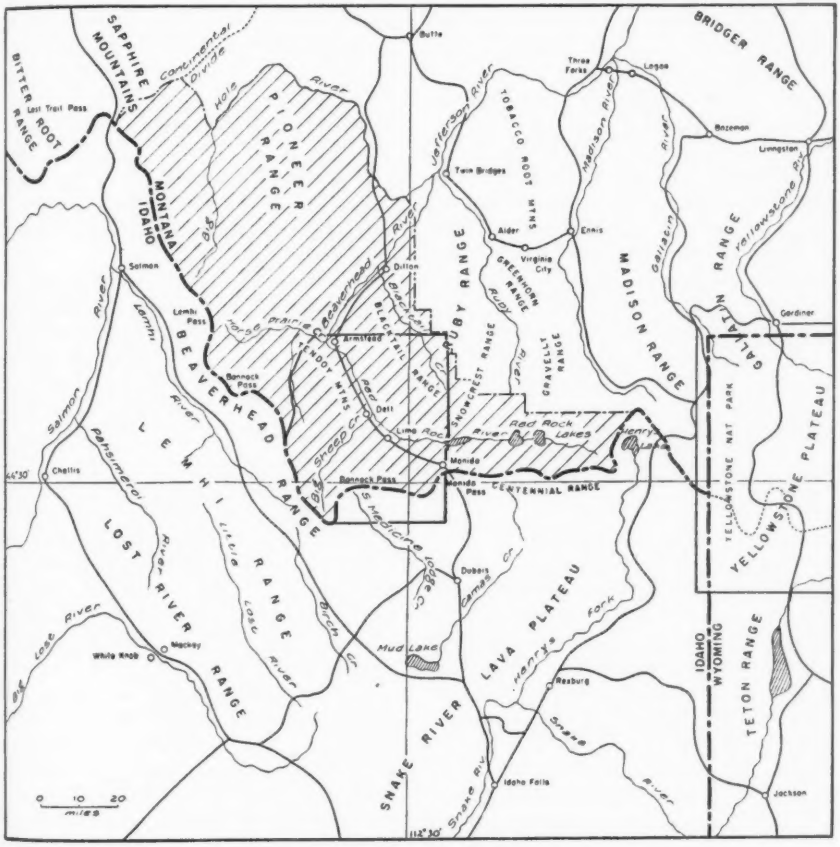

FIGURE 1.—LOCATION MAP
Cross-hatched area is Beaverhead County, Montana.

An unpublished reconnaissance map of southwestern Montana, which includes a part of the Lima region, was compiled during the summer of 1946 by E. S. Perry and U. M. Sahinen of the Montana Bureau of Mines and Geology. Perry (1948, p. 18–20; 1949, p. 17) describes in detail the graphite deposit near Kate Creek and the gypsum deposits near the East Fork of Little Sheep Creek. Pardee (1950, p. 366–369; 377–386; 402–403) discusses several physiographic and structural features of the Lima region in a report on Cenozoic faulting in western Montana. Hibbard and Keenmon (1950) describe Miocene sediments and vertebrates from the eastern part of the region, and Klepper (1951) included this same area in a general reconnaissance of parts of Beaverhead and Madison counties, Montana.

Of the metamorphic and sedimentary units in southwestern Montana that have recently been investigated in detail, the pre-Beltian is

discussed by Heinrich (1948; 1949; 1950; 1953), the Paleozoic by Sloss and Moritz (1951), the Permian by McKelvey (1949, p. 273–274), the Triassic and Jurassic by Moritz (1951), and the Paleocene by Lowell and Klepper (1953). All these reports include references to localities in the Lima region.

Present Study

This paper is based on investigations conducted individually by the writers during the summers of 1947 and 1948 in three areas within the Lima region, and presented in theses written for the degree of Doctor of Philosophy at the University of Michigan.

Mapping was done on aerial photographs on a scale of approximately 1:20,000. The Forest Service maps of Beaverhead, Targhee, and Salmon National forests, which indicate the land divisions, were used as controls. The boundaries of the three original areas are indicated on the index map (Plate 1).

Acknowledgments

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GEOGRAPHY

Topographic Features

The main topographic prominences are the Beaverhead, Tendoy, Blacktail, and Snowcrest ranges (Fig. 2). The Beaverhead Range, along the western boundary of the region, trends northwest. Its narrow crest line is between 9000 and 10,000 feet above sea level, but several peaks rise above 10,000 feet. The most rugged relief is within this range. Southeastward the Beaverhead Range plunges under the lavas of the Snake River plains, outside the limits of the Lima region.

An almost continuous intermontane depression, made up of South Medicine Lodge, Big Sheep Creek (or Nicholia), and North Medicine Lodge basins, separates these mountains from the Tendoy Range farther east. The Tendoy Range parallels the Beaverhead Range and also plunges and disappears southeastward. Except for several isolated peaks, its crest is gently rolling. The northwestern limit of the Tendoy Range is at the broad east-west valley of Horse Prairie Creek. Muddy Creek Basin is an intramontane depression in this range. To the east, Red Rock Basin parallels the Tendoy Range. It averages about 5 miles in width and has a mean altitude of approximately 5700 feet.

East of Red Rock Basin the trends of the mountains are diverse. The Snowcrest Range, in the eastern portion of the region, trends approximately northeast. South of it is the Lima Reservoir Basin, of which the western extremity lies in the Lima region. The Blacktail Range, in the northern part of the region, trends approximately northwest. Its gently rolling summit rises more than 9000 feet above sea level and more than 3000 feet above the broad, alluviated valley of Blacktail Creek which bounds the range to the northeast. Red Rock "Mountains" is a series of hills between Red Rock Basin and the Blacktail Range. The northeastern corner of the region covers a small portion of the north-trending Ruby Range.

Drainage

Blacktail Creek and its tributaries drain the northeast sector of the region. The central and

eastern portions are drained by the Red Rock River and its tributaries: Sage, Junction, Big Sheep and Little Sheep creeks. Big Sheep Creek follows a canyon which cuts through the entire width of the Tendoy Range. North Medicine Lodge Creek, along the west margin, empties northward into Horse Prairie Creek, which joins the Red Rock River near Armstead. The confluence is known as Beaverhead River; it meets Blacktail Creek north of the region to form the Jefferson River, one of the major streams of the Missouri River Basin.

The Continental Divide passes through the region, following the Idaho State line. Consequently, the southern part of the area, drained by South Medicine Lodge Creek and its tributaries, discharges via the Snake River to the Pacific. It occupies a canyon in the basin between the Tendoy and Beaverhead ranges.

FIGURE 2.—GEOMORPHIC DIAGRAM

Climate and Vegetation

The Lima region has a semi-arid, continental climate. In the main basins annual rainfall averages 9 inches, which is the lowest average in the northern Rocky Mountains. Mean temperatures are 18° F. during winter and 60° F. during summer. In the mountains the precipitation is higher, and the temperatures are lower, so that isolated snow banks persist through most of the summer.

Vegetation consists mainly of sagebrush and sparsely distributed conifers; the latter are re-

stricted to mountain slopes and are most abundant on northern exposures. Grass, and cottonwood and willow trees occur in main valleys and basins, where they are confined to the banks of streams.

Accessibility

U. S. Highway 91, connecting Idaho Falls, Idaho, with Butte, Montana, passes through Red Rock Basin in the center of the area. The Union Pacific Railroad parallels this highway. Gravel and improved dirt roads branch off from the highway bost east- and westward, following basins and canyons, and permitting access by automobile to, and partly into, the mountain ranges.

Metamorphic Rocks

Introduction

The metamorphic rocks of southwestern Montana are pre-Beltian and can be divided into three major units: Pony group, Cherry Creek group, and Dillon granite gneiss. The Pony group is dominantly orthogneisses and schists derived from igneous rocks. The Cherry Creek group contains chiefly marbles, schists, and metaquartzites. The youngest metamorphic unit, the Dillon granite gneiss, intrudes both Pony and Cherry Creek rocks.

Pre-Beltian System

Pony group.—The rocks of the Pony group are biotite gneisses, hornblende gneisses, and amphibolites. At least two types of hornblende gneisses seem to be present: (1) those containing appreciable feldspar; (2) those containing little or no feldspar.

The first type commonly exhibits gneissic banding in hand specimen, although the banding is not so conspicuous as in the acidic gneisses of the Dillon granite gneiss. Quartz is absent except in veinlets along which it was introduced later. Green hornblende is the most prominent dark mineral; quantitatively it is about as important as feldspar (andesine). Pyroxene (commonly hypersthene, but also augite and diopside) is present in varying amounts. The rock is the metamorphic equivalent of a diorite.

The second type of hornblende gneiss, with little or no feldspar, does not exhibit any banding in the hand specimen, in which the coarse amphibole is readily recognizable. The little feldspar present is altered almost beyond recognition. Epidote and smaller amounts of zoisite are prominent constituents. The pyroxene is diallage, characteristic of many gabbros and related rocks.

Some of the hornblende gneisses are cut by thin stringers of acid plagioclase or potash feldspar with or without quartz, commonly parallel to the lineation. Rocks containing these stringers are believed to represent hybrids produced during the later intrusion of granitic material of the Dillon pluton into the older hornblende gneisses. The hornblende gneisses grade into amphibolites which consist almost entirely of green hornblende in a granoblastic arrangement.

The gneisses and amphibolites occur in the core of the Armstead anticline, in the Kate Creek area, in the southeastern end of the Blacktail Range, and in the extreme southeastern end of the Ruby Range. Some of them probably represent metamorphosed sills and lenses of igneous rock intrusive in the Pony as well as the Cherry Creek (Heinrich, 1949, p. 23, 33). The absence of well-defined metasedimentary units over large areas is characteristic of the Pony group. Minor amounts of possibly sedimentary materials, represented by quartz-mica schists, were found in the eastern part of the area, but most of the Pony is clearly igneous. In small areas it is difficult to distinguish the Pony from the Cherry Creek group.

The stratigraphic relations between the Pony and the Cherry Creek are not clear in the Lima region. Farther east, Tansley *et al.* (1933, p. 10) found boulders of Pony gneiss in a schist at the base of the Cherry Creek. This relationship would suggest two periods of pre-Beltian metamorphism, one before and one after the Cherry Creek group. Heinrich (1953, p. 1432) considers the Pony to be older than the Cherry Creek.

Cherry Creek group.—In the Lima region gneisses, schists, marbles, and metaquartzites make up the Cherry Creek group.

Near the base of the group is a biotite-garnet gneiss, in which large feldspar crystals and almost black dots of garnet crystals are conspicuous. The feldspar is orthoclase and is considerably less

abundant than quartz. Green and brown biotite are present. The garnet has a poikiloblastic texture. Rocks containing abundant almandite (44%), no feldspar, but hypersthene (37%), quartz (17%), and accessory minerals (2%) are exposed in the Tendoy Range, section 16, T. 15 S., R. 10 W., Montana.

The most conspicuous type of schist in the Cherry Creek group is a graphite schist, in the lower part of the series exposed along Kate Creek in the Tendoy Range. The rock is gray, owing to small crystals of white-weathered feldspar and flakes of graphite. Schistosity is well developed. The feldspar may be orthoclase but is so altered that identification is ambiguous. Quartz approximately equals feldspar. The graphite occurs in slender needles which form bundles and long, streaky masses. The rock contains appreciable small rutile crystals, some of which are partially coated with leucoxene. Rutile is a characteristic associate mineral of graphite in metamorphic sediments of partially organic origin.

The upper part of the Cherry Creek group consists mainly of marbles and quartzites. The former are in part pure carbonate marbles and in part silicate marbles. The carbonate is mostly calcite; dolomite is rare or absent in 8 out of 10 specimens. The silicate marbles contain abundant serpentine, consisting mostly of irregularly oriented chrysotile fibers. In several specimens antigorite, or a mixture of chrysotile and antigorite, was found. Pyroxenes, amphiboles, micas, and chlorites are accessory. The pyroxenes are diopside and more rarely augite. The only amphibole found was tremolite. Most of the mica is phlogopite; some of it is intergrown in crystallographic continuity with green biotite and some with pennine. Some specimens contain a large amount of bluish crystalline quartz and of brown iron ore in alternating laminae. Brown stains of limonite are common.

The quartzites are less abundant than the marbles. A pale to vivid emerald-green quartzite is part of the Cherry Creek group as exposed in the Kate Creek area. The color seems to be due to the presence of a green mica. Heinrich (1950, p. 8) believes that this green quartzite is diagnostic of Cherry Creek rocks.

Cherry Creek rocks are exposed in the core of the anticline northwest of Armstead, along Kate Creek, and in the mass of Dillon granite gneiss between Nicholia Creek and Little Deadman Creek. The limited exposures preclude accurate estimate of the thickness.

The lithology of the series indicates that most of it is of sedimentary origin. The rocks show the influence of both regional and contact metamorphism. Regional metamorphism of impure dolomitic limestones produced the tremolite, diopside, and phlogopite in the Cherry Creek calcite marbles. Locally, purer limestones and dolomites were altered to pure carbonate marbles. The serpentine probably was produced through reaction of thermal solutions, rising from the invading Dillon batholith, with the tremolite and diopside. This type of replacement was observed in several specimens. Post-magmatic solutions probably introduced the silica found in localized veins and pockets in some of the marbles.

The Cherry Creek quartzites were evidently produced by regional metamorphism of sandy sediments, whereas the source material of the graphite and the biotite-garnet rocks was argillaceous and locally rich in carbonaceous matter.

Dillon granite gneiss.—The name Dillon granite gneiss has been introduced by Heinrich (1953, p. 1432) to replace the term Blacktail previously used for the same unit (Heinrich, 1950, p. 10), as the latter term is preoccupied (Wilmarth, 1938, p. 208). The Dillon granite gneiss is a plutonic complex of granite gneisses, pegmatites, aplites, quartz masses, and hybrid gneisses, characterized by a topography of rounded slopes covered with reddish soil and supporting meager vegetation of grasses and sagebrush.

The most common rock type is a reddish brown feldspar-quartz gneiss in which bands of pink feldspar and quartz alternate with dark bands of mafic minerals.

Quartz is universally present, and several types of feldspar (orthoclase, microcline, plagioclase) occur. The amount of plagioclase (albite to oligoclase) varies greatly. Biotite is common but in the coarser-grained parts of the gneiss dark minerals are very scarce. Garnet occurs locally but seems to be abundant in the rocks north of Armstead. Green hornblende is less common. Apatite and zircon are important accessory minerals; the latter is present throughout in large rounded grains.

The most conspicuous features in almost all of the Dillon granite gneiss are (1) cataclastic structures in feldspar and quartz crystals, commonly accompanied by strain shadows in the quartz; (2) perthitic, myrmekitic, and micropegmatitic intergrowths; (3) reaction rims around potash

feldspar and plagioclase where these are in contact.

Some coarse-grained rocks in the Dillon complex represent pegmatite dikes, consisting mostly of large crystals of pink, coarsely twinned feldspar (microcline or orthoclase) in graphic intergrowth with quartz. Perthitic intergrowths of orthoclase with albite are also common. Reaction rims of secondary quartz are visible around some of the quartz grains that are in contact with the microcline. The feldspars and quartz compose virtually the entire rock, and both show cataclastic structures. Specimens containing significant plagioclase (albite-oligoclase) may represent a different zone of the pegmatite. Metamorphic structures, mineralogical similarity to the normal, finer-grained Dillon granite gneiss, and local gradations into the latter indicate these pegmatites are Precambrian. A few pegmatites, however, show no metamorphism and could be Laramide.

Fine-grained meta-aplites with anhedral crystals are exposed at several places in the Dillon complex. The main minerals are quartz and feldspar (orthoclase and oligoclase). Perthitic intergrowths are particularly conspicuous. In addition quartz veins and irregular masses are common. Some may be younger than the Precambrian rocks.

Several types of hybrid rocks are present in the Dillon granite gneiss complex. Some are epidote-rich gneisses in which the abundance of quartz and the presence of microcline, orthoclase, oligoclase, and perthitic intergrowths suggest close connection to the ordinary Dillon granite gneiss. The local abundance of epidote and zoisite suggests that these rocks may have originated through the interaction of the Dillon granite with the older, more basic gneisses of the Pony group. However, some of the epidote gneisses probably are the products of retrogressive metamorphism of ordinary granite gneiss which caused saussuritization of the plagioclase.

The Dillon granite gneiss complex is the most widespread of the pre-Beltian metamorphic rocks. It outcrops over large areas of the Blacktail and Tendoy ranges and of the mountains north of Armstead.

The Dillon batholith intruded the Cherry Creek group during the post-Cherry Creek orogeny, which metamorphosed the impure Cherry Creek dolomites to silicate marbles. Volatile emanations of the granitic intrusion probably were responsible for the serpentinization of these silicate marbles during the later stages of the orogeny. Late orogenic pulsations are responsible for the foliation of the granite gneiss. Late phases of the intrusion may have been post-tectonic, introducing some of the nonfoliated and pegmatitic varieties of the complex, as well as the quartz veins in the Cherry Creek marbles. Apparently, the magma had a relatively high water content. A comparatively low viscosity is also suggested by *lit-par-lit* injections into older gneisses, notably in the Armstead area.

Sedimentary Rocks

Introduction

Stratigraphic column.—Sedimentary rocks in the region represent all systems from Beltian through Quaternary, except the Silurian. The Paleozoic strata thicken westward to a maximum thickness in excess of 10,000 feet. Carbonate sediments make up much of these Paleozoic rocks. The dominantly clastic Mesozoic strata attain a maximum thickness of more than 7500 feet in the southern part of the region. Disconformities exist between many of the formations, but angular unconformities only between the pre-Beltian and younger rocks, and at the bottom and top of the Paleocene.

The Mesozoic sediments are present only east of a line approximately along the crest of the Tendoy Range. Paleozoic and Precambrian rocks occur on both sides of this line. This dividing line is part of a regional western limit of Mesozoic rocks in the northern Rocky Mountains, noted by Ross (1947, p. 1126).

The sediments demonstrate a gradual transition from chiefly marine beds in the systems of the Beltian, Paleozoic, and lower Mesozoic to dominantly terrestrial deposits in the systems of the upper Mesozoic and Cenozoic. The most important features characterizing each litho-unit are summarized in Table 1. Detailed descriptions of the Paleozoic, Mesozoic, and Paleocene sedimentary rocks in this part of southwestern Montana have been given by Sloss and Moritz (1951; Moritz, 1951) and by Lowell and Klepper (1953; Klepper, 1951).

Paleotectonic control.—The sedimentary rocks in southwestern Montana and adjacent parts of Idaho clearly display the pattern of the tectonic framework that controlled their deposition. With respect to thickness as well as

lithology some sedimentary units, such as the Cambrian System, the Devonian Three Forks formation, the Mississippian Big Snowy (?) group, and the lower Mesozoic formations have most of the characteristics of deposition on a tectonic shelf. Others, notably the Beltian System and the Mississippian Milligen and Brazer formations, evidently were laid down in a geosyncline.

The transitions between these two main facies are horizontal as well as vertical and may be observed within the Lima region. The horizontal transitions reflect the critical tectonic position of this area between the axis of the Rocky Mountain miogeosyncline and the stable mass of the craton to the east and northeast. This pattern was established in Precambrian time and persisted through part or most of the Mesozoic.

The vertical transitions reflect fluctuations of the geosynclinal margin and the resulting cyclic deposition demonstrated for this part of the Cordilleran region by Sloss (1950, p. 450–451). Not all these cycles are traceable within one relatively small area, but Sloss's third cycle (Kaskaskia sequence), in particular stages two, three, and four (inceasing definition and final culmination of differentiation of tectonic framework, followed by general uplift and erosion), are excellently recorded in the sediments of the lower Carboniferous. Both facies of the sharply differentiated framework have been brought even closer by Laramide thrusting, so that they are now found within the limits of the Lima region.

Paleogeographic control.—The materials for most of the marine sediments were derived from low-lying land areas in the east or northeast and were deposited in shallow seas. Notable exceptions, however, are the shales of the Mississippian Milligen formation and the sandstones of the Belt group, which probably were derived from the west or southwest. Some sedimentary deposits such as the marine Jurassic strata may have received contributions from both directions. At frequent intervals, notably during the Ordovician, Silurian, Early Devonian, Late Triassic, and Early Jurassic, the area, or parts of it, were dry and became the site of nondeposition or erosion. The late Mesozoic terrestrial deposits probably were derived from mountainous areas in the west, and during the Cenozoic the sources of the sediments were largely local.

Beltian System

The Belt group, uniform throughout, consists predominantly of maroon, green, and light-gray graywackes (Pl. 2, fig. 1) with minor conglomerate and an occasional lens of brown chert. The Belt rocks are confined to the western part of the region. They are well exposed on Island Butte, in the northern part of Big Sheep Creek Basin, and along Trail Creek in the Beaverhead Range. One questionable outcrop occurs in the Tendoy Range, east of North Medicine Lodge Creek. No Beltian occurs east of the Tendoy Range. To what extent this is due to nondeposition or to post-Beltian erosion is undetermined. The thickness is unknown, as the base of the series is nowhere exposed, but the minimum thickness is 1000 feet. The total thickness probably is much greater.

The only exposure of presumed Beltian rocks directly overlying pre-Beltian rocks is near Warm Springs Creek in the Tendoy Range. The contact is not exposed, but the difference in the degree of metamorphism indicates that it is a profound unconformity. In the absence of Cambrian strata in the area of Belt distribution the Belt rocks are overlain by white quartzites tentatively assigned to the Ordovician Kinnikinic (?) formation. The nature of this contact is not evident in the field, but if the correlation is correct it represents an important disconformity. The maroon quartzites below the Kinnikinic (?) were considered Beltian on the basis of lithologic similarity to rocks described as Belt in near-by areas and dissimilarity to the Cambrian Flathead quartzite. They may correlate with the Spokane formation of the series. Although in the type area the Spokane rocks are dominantly shaly, sandstones become more important to the west, especially near the top (Emmons and Calkins, 1913, p. 35).

Cambrian System

The Cambrian formations include, in order of deposition, the Middle Cambrian Flathead

TABLE 1.—SEDIMENTARY ROCKS IN THE LIMA REGION

Name and age of rock unit	Composition	Colors	Textures	Bedding and structures	Fossils	Exposures	Approximate thickness (Feet)
Quaternary							
Recent	Floodplain clays, silts; terrace gravels, fanglomerates, and landslide deposits		Gravels well-rounded				
Pleistocene							
Pinedale (?) drift	Boulderclay						
Bull Lake (?) drift	Boulderclay						
Tertiary							
Pliocene (?)	Stream gravel		Well-rounded to subrounded pebbles, cobbles			Strewn across intermediate erosion surface in Beaverhead and Tendoy ranges	
Miocene							
W. & S. of Highway 91: Medicine Lodge beds	Shales, bentonitic clays, sandstones, pebble conglomerates, fanglomerates; some lignite; thick, vuggy freshwater limestone at top	White, gray, yellow, violet brown, brown	Sandgrains poorly sorted, fine to coarse, subrounded	Thin- to medium-bedded, crossbedded, jointed		Limestone supports cliffs in South Medicine Lodge Basin; other units poorly exposed along Nicholia Creek in Big Sheep Creek Basin	5000+
E. & N. of Highway 91: Blacktail Deer Creek fm.	Clays, feldspathic sandstones, fine conglomerates, tuff; thick, vuggy travertine limestone; basal pebble conglomerate and limestone breccia; interbedded basalt flows; grains, pebbles and thin beds of chert; some lignite	Light gray and tan; some weathering purplish	Poorly sorted	Thin- to medium-bedded	*Caenopus* sp. *Arretotherium acridens* Douglass *Monosaulax herperus* (Douglass) (Coll. by Douglass, 1902, p. 239) *Merycodus* sp. Horse fragments (Coll. by J. Leroy Kay, pers. comm.) *Promerycochoerus barbouri* Schultz and Falkenbach *Gregorymys montanensis* Hibbard and Keenmon	Small outcrops; limestone locally supports cliffs; type section near center, sec. 8, T. 11 S., R. 6 W., Montana, along Blacktail Cr.	2000+
Oligocene							
W. & S. of Highway 91: Muddy Creek beds	Tuff, with some shales, conglomerates and fresh-water limestone; basal pebble conglomerate and sandstone in Muddy Hole Basin; some bituminous shales	Mostly white, light gray and buff; some pink, red, brown, and black			Sequoia cones (Coll. by Bowen, 1917, p. 317) *Planorbis* cf. *utahensis* Meek *Lymnaea* cf. *sinuscula* White *Oreohelix* sp.	Poorly exposed in Muddy Creek and Muddy Hole basins	1000+
E. & N. of Highway 91: Cook Ranch fm.	Clays, tuff, some sandstones	Mostly white and		Local channel de-	*Peratherium* sp.	Poorly exposed; type	125–1000+

E. & N. of Highway 91 — Cook Ranch fm.	Clays, tuff, some sandstones and conglomerates; grades northward into tuffs and acidic breccia	Mostly white and light gray		Local channel deposits, cross-bedding	*Peratherium* sp. *Heliscomis gregoryi* Wood *Eumys* sp. *Hyracodon* sp. *Palaeolagus haydeni* Leidy *Leptomerix evansi* Leidy (Coll. by Wood, 1934, p. 254) *Daphoenus dodgei* Scott *Caenopus* (or *Trigonias*) sp. *Mesohippus* sp. *Merycoidodon* sp. Oreodons (Coll. by J. Leroy Kay, pers. comm.)	Poorly exposed; type section re-established in secs. 27, 34, T. 12 S., R. 8 W., Montana, in Sage Creek valley	125–1000+
Eocene							
Sage Creek fm.	Sandstones, conglomerates, clays, tuff; basal basalt	Gray and greenish	Sand grains dominantly fine	Local channel deposits, cross-bedding	*Helaletes* (?) sp. *Hyrachyus douglassi* Wood *Chasmotheroides* cf. *intermedius* *Amynodoni* (?) sp. (Coll. by Wood, 1934, p. 249–251) *Desmatotherium* sp. *Diplophodon* (?) sp. *Protoreodon medius* Peterson (Coll. by J. Leroy Kay, pers. comm.)	Nonresistant; "badland" topography; type section re-established in S.W. ¼ sec. 28, T. 12 S., R. 8 W., Montana, in Sage Creek valley	300–400
Angular unconformity							
Paleocene							
Beaverhead fm.	Pebble and boulder conglomerate; some "house-size" boulders; sparse sandy and silty matrix; mostly quartzite, locally limestone and chert pebbles; toward southeast thick "salt-and-pepper" sandstones occur near base, finally constitute much of formation; thick fresh-water limestone unit intercalated in McKnight Canyon, Tendoy Range	Red, orange, and gray	Poorly consolidated, ill-sorted, well rounded to angular (limestone breccia); limestone unit oölitic, with algal concretions	Poorly bedded to well bedded	*Lioplacodes* cf. *tenuicarinata* (Meek and Hayden) *Lioplacodes* cf. *limnaeiformis* (Meek and Hayden) *Amnicola* sp. *Hydrobia* sp. *Pleurocera* cf. *warrenanum* (Meek and Hayden) *Goniobasis* (?) sp. *Pleurolimnaea* sp. *Physa* sp. *Holospira* cf. *grangeri* Cockerell *Oreohelix* sp. (Coll. by Lowell and Klepper, 1953, p. 240)	Supports rolling hills, locally cliffs; type section near mouth of McKnight Canyon, Tendoy Range, T. 12 S., R. 10 W.; other exposures along U. S. Highway 91	9700+
—Angular unconformity							

TABLE 1—*Continued*

Name and age of rock unit	Composition	Colors	Textures	Bedding and structures	Fossils	Exposures	Approximate thickness (Feet)
Cretaceous							
Upper (?) Cretaceous							
Aspen fm.	Shale, siltstone, "salt-and-pepper" sandstone, locally conglomeratic; some thin lithographic limestones; interbedded bentonite and porcellanite; volcanic breccia (quartz, sanidine, plagioclase, carbon) in upper portion	Dominantly light gray; variegated weathering colors ("paint pots")	Poorly sorted sand grains		*Cissites* (or *Platanus*) *affinis* *Betula beatriciana* Bone fragments of crocodile or small dinosaur	Poorly exposed in gently rolling topography; local "badlands"; sandstones support ledges; best outcrops near Sawmill Creek, Tendoy Range	0–3500
Lower Cretaceous							
Kootenai fm.	"Salt-and-pepper" sandstone, partly conglomeratic, with interbedded shales and some fresh-water gastropod limestones; basal cherty conglomerate; coquina at top	Gray, brown, and reddish; limestones white	Sand grains poorly sorted, subangular, coarse near base, fine in upper part		*Viviparus montanensis* Stanton *Campeloma harlowtonensis* Stanton *Unio douglassi* *Unio* cf. *hamili* McLearn	Sandstones support ledges; partly talus-covered; best section in area of Sawmill Creek, Tendoy Range	900–1500
Jurassic							
Upper Jurassic							
Morrison fm.	Shale, siltstone, and minor "salt-and-pepper" sandstone; locally carbonaceous shales and coal seams at top	Red, tan, and greenish gray	Sand grains fine to medium			Largely concealed by red or brown soil; poorly exposed near Little Water Cr. and headwaters of Little Sheep Cr. (Middle Fork), Tendoy Range	0–330
Swift fm.	Sandstone, glauconitic and calcareous, with some chert; some sandy shale	Greenish	Sand grains medium to coarse			No natural exposures	0–10
Rierdon fm.	Limestone; some calcareous shale near top	Gray, buff, and tan; shales greenish gray and tan; weathers white	Oölitic	Thin- to medium-bedded		Basal unit supports prominent ledges along Deep Cr. and Little Water Cr., Tendoy Range	0–170
Sawtooth fm.	Calcareous shale and siltstone, with minor sandstone; some silty limestone	Light gray and tan; weathers brown, white,			*Ostrea strigilecula* *Pleuromya weberensis* *Camptonectes* sp.	Mostly concealed by light-gray to yellowish-brown soil;	0–200

Sawtooth fm.	Calcareous shale and siltstone, with minor sandstone; some silty limestone	Light gray and tan; weathers brown, white, mottled			*Ostrea strigilecula* *Pleuromya weberensis* *Camptonectes* sp. *Lima* sp. *Gryphea planoconvexa*	Mostly concealed by light-gray to yellowish-brown soil; poorly exposed at head of Little Sheep Cr. (Middle Fork) and in Little Water Canyon, Tendoy Range	0–200
Triassic Lower Triassic Thaynes fm.	Silty limestones, locally cherty, at top and bottom; middle unit of calcareous siltstone and sandstone	Gray and tan; weathers pink, light gray, brown	Fine-grained	Thin- to medium-bedded	*Pugnoides triassicus* *Pentacrinus* sp.	Resistant units support ledges in Tendoy (Little Water Cr., Little Sheep Cr.), Snowcrest (W. Fork, Blacktail Cr.), and Blacktail ranges	630–950
Woodside fm.	Calcareous shale and siltstone; minor shaly limestone	Red, maroon, and gray				Concealed by maroon and reddish soils beneath smooth slopes and strike valleys; poorly exposed along Little Water and Little Sheep (W. Fork) creeks, Tendoy Range	0–190
Dinwoody fm.	Calcareous siltstone and shale; some silty limestone	Gray, brownish gray, and reddish brown; weathers chocolate brown		Thin-bedded; jointing produces blocky appearance	*Anadontophera fassaensis* (Wissman) *Anadontophera canadensis* Catullo *Anadontophera* sp. *Claraia mulleri* *Claraia stachei* *Gervillia* cf. *explorecta* *Eumorphotis multiformis* Bittner *Eumorphotis* (?) sp. *Lima* cf. *striata* Schlotheim *Monotis* (?) sp. *Myalina* sp. *Myalina putiatinensis* Kiparisova *Pseudomonotis* (?) sp. *Ambocoella* (?) sp. *Lingula borealis* Bittner *Clypites* sp. *Meekoceras* sp. *Ophiceras dubium* (?) Spath *Ophiceras* cf. *subplatyspira* Spath *Ophiceras subdemissum* (?) Spath	Locally supports ledges; mostly concealed by brown soil; best exposures along Blacktail Cr. (W. Fork), Snowcrest Range, and along Little Water, Little Sheep (Middle Fork, headwaters) and Deep creeks, Tendoy Range	180–870

TABLE 1—*Continued*

Name and age of rock unit	Composition	Colors	Textures	Bedding and structures	Fossils	Exposures	Approximate thickness (Feet)
Permian							
Phosphoria fm.	Alternating units of limestone and dolomite; phosphatic and kerogen shales with phosphate beds; chert beds; and sandstone	Light to dark gray, black and brown; bluish bloom on phosphatic shale; red unit near base	Phosphate oölitic	Thin-bedded; shales fissile	*Dictyoclostus* sp. *Productus* sp. *Pustula nevadensis* *Echinoconchus nevadensis* (Meek) *Composita parva* Branson *Composita plana* Branson	Largely concealed; partial exposures in canyons of Big Sheep C.r and Little Sheep Cr. (Middle Fork, headwaters), Tendoy Range; also along W. Fork of Blacktail Creek, and at Wadham's Springs, Snowcrest Range	360–830
Pennsylvanian							
Quadrant fm.	Orthoquartzite; cement siliceous and calcareous; minor cherty dolomite, mostly near top	White to buff; weathers brown, red, and purple	Fine- to medium-grained; tight to friable	Thick-bedded to massive; cross-bedding common	Fusilinids	Supports cliffs and high peaks, skirted by talus; well exposed in Big Sheep Creek canyon, Tendoy Range, and near Red Rock River, Snowcrest Range	1000–2625
Amsden fm.	Limestone with some limestone conglomerate; shale, partly partly sandy, calcareous; basal sandstone	Gray-to-buff, red		Thin-bedded		Largely concealed by reddish- or yellowish-brown soil and Quadrant talus; partial exposures in Big Sheep Creek canyon, Tendoy Range, and near Red Rock River, Snowcrest Range	90–200

Mississippian Upper Mississippian								
Tectonic shelf suite "Montana facies"	Amsden fm. Big Snowy group	Limestones and shales in upper part; shale, siltstone, sandstone, and gypsum in lower part	Upper part black and gray; lower part reddish	Limestones fine-fragmental to dense	Thin-bedded; shales fissile	*Camarotoechia* (?) sp. *Camarotoechia metallica* White *Cliothyria crassicardinalis* White *Dictyoclostus* (?) sp. *Dictyoclostus burlingtonensis* (Hall) *Leptaena rhomboidalis* (Wilckens) *Linoproductus ovatus* (Hall) *Leiorhynchus* sp. *Martinia* sp. *Orthotetes* sp. *Reticularia* (?) sp. *Schizophoria* sp. *Seminula* sp. *Seminula subtilita* (?) Girty *Spirifer* sp. *Spirifer centronatus* Winchell *Spirifer* aff. *keokuk* Hall *Spirifer striatus madisonensis* Girty *Syringothyris* sp. *Myalina* sp. *Caninia* sp. *Triplophyllites* sp. *Fenestrellina* sp.	Largely concealed; partial exposures in Big Sheep Creek canyon, Tendoy Range, and along Blacktail Creek (Middle Fork), Snowcrest Range	0–500+?
Geosynclinal suite "Idaho facies"	Brazer fm.	Limestone, partly coralline; chert layers and nodules; minor shale and limestone breccia	Dark gray and black; weathers gray, locally pinkish	Fine-grained to dense	Thin-bedded to massive; partly laminated and cross-bedded	*Caninia* sp. *Clysiophyllum* sp. *Lophophyllidium* sp. *Syringopora surcularia* *Lithostrotion whitneyi*	Supports cliffs, high ridges, and peaks, skirted by talus; well exposed at head of Tex and Nicholia creeks, Beaverhead Range	0–2000+
	Milligen fm.	Shale, partly calcareous, with minor limestone	Black; weathers gray		Fissile		Forms smooth slopes, locally supports ledges; exposed at head of Nicholia Cr., Beaverhead Range	0–2000+

TABLE 1—*Continued*

Name and age of rock unit	Composition	Colors	Textures	Bedding and structures	Fossils	Exposures	Approximate thickness (Feet)
Lower Mississippian							
Mission Canyon fm.	Limestone, with chert layers and nodules; local limestone breccia	Dark to bluish gray; weathers light to medium gray	Fine-grained	Largely massive to thick-bedded; large solution cavities near top; some cross-bedding	*Chonetes ornatus* (?) Shumard *Dictyoclostus burlingtonensis* (Hall) *Leptaena rhomboidalis* (Wilckens) *Linoproductus ovatus* (Hall) *Linoproductus* sp. *Orthotetes inequalis* Hall *Orthotetes inflatus* Hall and Whitfield *Productella cooperensis* (?) (Swallow) *Seminula* (?) sp. *Spirifer centronatus* Winchell *Spirifer striatus madisonensis* Girty *Aviculopecten* sp. *Syringopora surcularia* *Triplophyllites* sp. *Strapoporollus* sp. *Strapoporollus utahensis* Hall and Whitfield *Platycrinus* sp.	Supports prominent cliffs; well exposed in Big Sheep Creek canyon, Tendoy Range; Ashbough canyon, Blacktail Range; and along W. Fork of Blacktail Cr., Snowcrest Range	700–1700+
Lodgepole fm.	Limestone, partly silty, cherty; calcareous shale	Dark-gray to bluish-gray	Fine-grained	Thin-bedded, even-bedded	*Dictyoclostus burlingtonensis* (Hall) *Martinia rostrata* Girty *Orthotetes inequalis* Hall *Orthotetes inflatus* Hall and Whitfield *Spirifer centronatus* Winchell *Platyostoma* sp. *Platyceras* sp.	Locally supports ledges; well exposed along Highway 91, 3 miles N. of Armstead; in Big Sheep Creek canyon, Tendoy Range, and near Ashbough Canyon, Blacktail Range	550–800
Devonian							
Upper Devonian							
Three Forks fm.	Calcareous shales and siltstones; shaly limestones	Brown, yellow, and reddish-gray; some black		Thin-bedded	*Chonetes* sp. *Cleiothyridina devonica* (Raymond) *Cyrtospirifer* (?) sp. *Leptaena* sp. *Leptaena rhomboidalis* (Wilckens) *Spirifer* sp. *Aulopora* sp. *Fenestella* sp. *Penniretopora* sp. *Brachymetopus* (?) sp.	Largely concealed by grayish-brown soil with limestone-shale float; partial exposures in Ashbough Canyon, Blacktail Range	115–600
Jefferson fm.	Dolomite, with shaly partings	Pinkish-gray,	Fine-grained	Massive to thin-bedded	*Leptaena rhombiodalis* (Wilckens)	Supports cliffs; well exposed in Ash-	0–750

Jefferson fm.	Dolomite, with shaly partings	Pinkish-gray, weathering mottled brown	Fine-grained	Massive to thin-bedded	*Leptaena rhombiodalis* (Wilckens)	Supports cliffs; well exposed in Ashbough Canyon, Blacktail Range	0–750
Ordovician							
Upper Ordovician							
Kinnikinic fm.	Orthoquartzite; local basal conglomerate	White and lavender	Grains fine (0.1-0.3 mm), equigranular, well rounded to subrounded, well sorted; hard and dense	Medium- to thick-bedded		Supports ledges; best exposure at head of Trail Creek, Beaverhead Range	0-800
Cambrian							
Upper Cambrian							
Pilgrim fm.	Dolomite, minor shale near base	Gray, weathering white; some tan	Saccharoidal	Massive		Supports cliffs; well exposed in Ashbough Canyon, Blacktail Range	0–225
Park fm.	Micaceous shale; minor dolomitic limestone and intraformational, edgewise limestone conglomerate	Greenish-gray; minor reddish-brown		Fissile		Largely concealed; poorly exposed in Ashbough Canyon, Blacktail Range	0–75
Meagher fm.	Dolomite, glauconitic near top; minor shale near top	Pink, tan, and reddish brown; weathering tan and buff	Fine-grained	Irregular bedding		Supports cliffs; well exposed in Ashbough Canyon, Blacktail Range	0–530
Wolsey fm.	Micaceous shale with calcareous nodules; some thin sandstone lenses	Grayish green	Sand grains fine	Fissile	Trilobite fragments	Largely concealed; poorly exposed in Ashbough Canyon, Blacktail Range	0–150
Flathead fm.	Orthoquartzite; scattered quartz pebbles; conglomerate beds near base; siliceous cement	Reddish brown	Medium- to coarse-grained, well rounded, well sorted	Medium-bedded; locally cross-bedded		Supports steep slopes; partly concealed by talus; exposed in Ashbough Canyon, Blacktail Range; at head of Kate Cr., Tendoy Range; and N. of Armstead	0–125

TABLE 1—*Concluded*

Name and age of rock unit	Composition	Colors	Textures	Bedding and structures	Fossils	Exposure	Approximate thickness (Feet)
Beltian Belt group	Graywacke (quartz dominant, 15–35% altered K feldspar and andesine, <5% zircon, magnetite, ilmenite, titanite, apatite, tourmaline, muscovite, biotite; matrix of sericite, chlorite, hematite, silica, is most abundant in greenish beds (approx. 25%); minor conglomerate and chert	Light gray, green, and maroon	Sand grains fine (0.1–0.3 mm), poorly sorted, poorly rounded; dense, hard	Thick-bedded, locally cross-bedded; laminated; penecontemporaneous slump structures common		Supports steep slopes, skirted by coarse talus; well exposed in Trail Creek Canyon, Beaverhead Range	0–1000+

sandstone, Wolsey shale, Meagher limestone, and Park shale and the Upper Cambrian Pilgrim limestone (or Hasmark formation according to Sloss and Moritz, 1951, p. 2144–2145).

The Cambrian strata total about 1100 feet in thickness in the eastern part of the area where they are exposed in the Blacktail Range (Pl. 3, fig. 1). Excellent outcrops occur in Ashbough Canyon. Here, the thicknesses of the individual formations are: Flathead (123 feet, measured), Wolsey (152 feet, measured), Meagher (528 feet, measured), Park (74 feet, measured), Pilgrim (225 feet, estimated). The Flathead and Wolsey formations have about the same thickness in the mountains north of Armstead where the Meagher has thinned to about 200 feet and the Park and Pilgrim formations are absent. Farther south, near Kate Creek in the Tendoy Range, only the Flathead and Meagher remain, each about 20 feet thick. No Cambrian formations occur in the southwestern part of the region.

Where the sequence is continuous the formations are conformable. The Flathead formation rests with distinct angular unconformity on the pre-Beltian complex. No Flathead was observed in contact with Belt quartzites, but the contact has been observed north of the Lima region (Winchell, 1914, p. 23). The Cambrian strata are overlain disconformably by the Devonian Jefferson formation.

Ordovician System

In the southwestern part of the area the Beltian rocks are overlain by strata lithologically similar to the Upper Ordovician Kinnikinic quartzite of central Idaho and so identified by Sloss and Moritz (1951, p. 2148). The formation supports conspicuous ledges in the Beaverhead and Tendoy ranges. Along Trail Creek its thickness was estimated to be 800 feet. From there it thickens rapidly to about 3000 feet in the geosynclinal region of central Idaho (Ross, 1947, p. 1095) but pinches out to the north and east in the Lima region.

The Kinnikinic formation is overlain by the Upper Devonian Threeforks formation.

Devonian System

The Devonian system in the Lima region comprises the Upper Devonian Jefferson dolomite and the overlying Three Forks shales and limestones. In the Blacktail Range the Jefferson formation is well exposed in Ashbough Canyon, where it is 150 feet thick. In the Armstead area more than 750 feet of strata was assigned to the Jefferson. The formation is present in the Tendoy Range near Kate Creek but is missing farther south. About 300 feet of Three Forks was measured near Armstead and in the Tendoy Range to the south. Westward it thickens to about 600 feet in the Beaverhead Range and eastward it thins to 115 feet in the Blacktail Range. The best fossil locality is in Ashbough Canyon.

Where both are present the Jefferson and Three Forks formations appear to be conformable, and the Devonian rocks rest on Upper or Middle Cambrian formations. In the southwestern part of the region the Three Forks is underlain by the Ordovician Kinnikinic formation. The Three Forks formation is overlain by limestones of the Madison group of Mississippian age with apparent conformity.

Mississippian System

In the Lima region the following Mississippian units were mapped: Madison group (Lodgepole and Mission Canyon limestones), Milligen shale, Brazer limestone, and Big Snowy (?) group. The Amsden formation may contain some beds of Mississippian age, but the greater part probably belongs to the Pennsylvanian. The identification of the sedimentary strata between the Amsden and Mission Canyon formations as the Big Snowy (?) group is uncertain and tentative; it has previously been made by Perry (1949, p. 17) and Sloss and Moritz (1951, p. 2158). Possibly the reddish shaly and silty gypsiferous beds near the base represent the Kibbey formation of the Big Snowy group. The medium gray to black shales interbedded between dark limestones in the higher part of the section are lithologically similar to the Heath formation.

The Mississippian units listed above are not in continuous sequence. The Brazer and Milligen formations are confined to the south-

western corner of the region where they are exposed in the Beaverhead Range along the Idaho State line. Incomplete sections measured in this vicinity indicate thicknesses around 2000 feet for both formations, but the actual thicknesses may be considerably greater. The Big Snowy (?) group is restricted to the Tendoy, Blacktail, and Snowcrest ranges. Owing to lack of adequate exposures the thickness of the group is unknown; at least several hundred feet appears to be present in the Tendoy Range. The Amsden formation, discussed below, is also restricted to the central, northern, and eastern parts of the region. The Mission Canyon limestones are distributed across the entire region (Pl. 3, fig. 2), but the Lodgepole formation (Pl. 4, fig. 1) has not been identified in the Beaverhead Range. The

TABLE 2.—PERMO-CARBONIFEROUS CORRELATION CHART
(With data from Sloss and Moritz, 1951, Fig. 9)

AGE		CENTRAL IDAHO	SOUTHWESTERN MONTANA			CENTRAL MONTANA
			SCHOLTEN		KEENMON	
		SLOSS & MORITZ	"IDAHO FACIES"	"MONT FACIES"	BLACKTAIL-SNOWCREST	SLOSS & MORITZ
PERMIAN		CASTO VOLCANICS	ABSENT	PHOSPHORIA FM	PHOSPHORIA FM.	ABSENT
PENNSYLVANIAN	VIRGIL	WOOD RIVER FM.	QUADRANT FM	QUADRANT FM	QUADRANT FM.	ABSENT
	MISSOURI					
	DES MOINES					
	MORROW		ABSENT ?	AMSDEN FM.	AMSDEN FM.	
MISSISSIPPIAN	CHESTER	BRAZER FM.	BRAZER FM.	BIG SNOWY GROUP	BIG SNOWY GROUP	LOWER AMSDEN FM BIG SNOWY GROUP: HEATH FM, OTTER FM, KIBBEY FM
	MERAMEC		MILLIGEN FM.	?	?	
	OSAGE	MILLIGEN FM.	MISSION CANYON FM	MADISON GROUP	MADISON GROUP MISSION CANYON FM	MADISON GROUP MISSION CANYON FM
	KINDERHOOK		LODGEPOLE FM.		LODGEPOLE FM	LODGEPOLE FM.

PLATE 2.—FEATURES OF SEDIMENTARY ROCKS

FIGURE 1.—SLUMP STRUCTURE IN BELT GRAYWACKE FROM TRAIL CREEK CANYON, BEAVERHEAD RANGE

FIGURE 2.—TYPICAL BEAVERHEAD CONGLOMERATE
Exposed at mouth of Big Sheep Creek, Tendoy Range.

PLATE 3.—PALEOZOIC CARBONATE ROCKS

FIGURE 1.—MEAGHER DOLOMITE (FOREGROUND) AND PILGRIM DOLOMITE (SKYLINE)
Exposed 1 mile above mouth of Asbough Canyon, Blacktail Range. View looking north. Photo by T. B. Piper.

FIGURE 2.—MISSION CANYON LIMESTONE
Exposed in west wall of Big Sheep Creek canyon, Tendoy Range, near Muddy Creek.

FIGURE 2

FIGURE 1

FEATURES OF SEDIMENTARY ROCKS

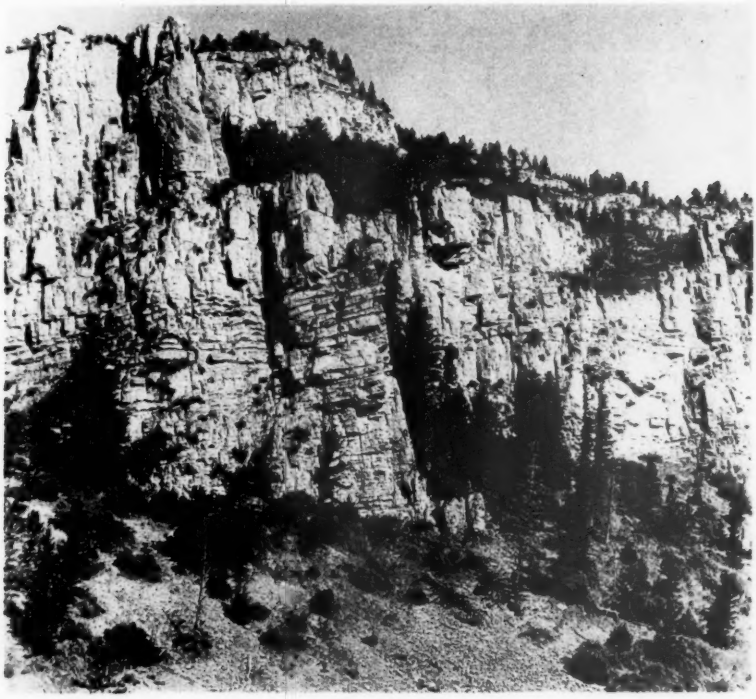

FIGURE 1

FIGURE 2

PALEOZOIC CARBONATE ROCKS

FIGURE 1

FIGURE 2

FIGURE 3

STRUCTURAL FEATURES

FIGURE 1

FIGURE 2

FIGURE 3

GEOMORPHIC FEATURES

thickness of the Madison group reaches 2500 feet in the Tendoy Range.

These relations reflect the sharp tectonic differentiation of the northern Rocky Mountain region during the lower Carboniferous. The Milligen and Brazer facies represent the Rocky Mountain miogeosycline whose axis lay to the west in central Idaho, whereas the approximately equivalent Big Snowy (?) and Amsden strata represent the eastern edge of a relatively more stable mass, the tectonic shelf of the adjacent craton. Not since Beltian time had truly geosynclinal conditions existed in the region. For convenience the geosynclinal suite may be called "Idaho facies", and the shelf suite "Montana facies" (Table 2). The presence of both facies within the region is partly due to considerable crustal shortening along the Medicine Lodge thrust fault during the Tertiary. The "Idaho facies" is confined to the thrust sheet which moved from southwest to northeast; the "Montana facies" occurs only in the autochthonous rocks. Similar observations were made by Sloss and Moritz (1951, p. 2153, 2160).

Although the Madison limestone belongs dominantly to the Lower Mississippian, the upper beds may be Upper Mississippian (Weller *et al.*, 1948, p. 139; Sloss and Moritz, 1951, p. 2157). The contacts between the Three Forks, Lodgepole, and Mission Canyon formations appear to be conformable. The Milligen shales are in fault contact with the Madison. In central Idaho they range from uppermost Devonian to middle Chesterian (Weller *et al.*, 1948, p. 140). In the Beaverhead Range, however, Madison limestone on top of Three Forks strata indicates that only the upper Milligen is represented here. Hence, these beds are Middle Mississippian and may be correlative with the uppermost part of the Madison group and with part or all of the Big Snowy (?) group in the "Montana facies".

The Brazer formation is chiefly Upper Mississippian, but the occurrence of *Lophophyllidium* may indicate that the upper beds are Pennsylvanian as the first occurrence of this genus is reported to be in the Pennsylvanian (Shimer and Shrock, 1944, p. 87). The contact between the Brazer and the underlying Milligen appears conformable. Near Morrison Lake Pennsylvanian Quadrant quartzites rest on the Brazer strata.

At the top of the Mission Canyon limestone solution cavities containing reddish and yellowish Big Snowy (?) material suggest a disconformable contact. The contact between the Big Snowy (?) group and the Amsden formation is gradational and appears conformable. The group is assigned to the Upper Mississip-

PLATE 4.—STRUCTURAL FEATURES

FIGURE 1.—DRAG FOLDS IN LODGEPOLE LIMESTONE

Exposed on northwest side of U. S. Highway 91, 3 miles north of Armstead. Photo by T. B. Piper.

FIGURE 2.—MEDICINE LODGE OVERTHRUST

Exposed on west side of Little Sheep Creek, near Two Spring Gulch, Tendoy Range. Light-gray Mission Canyon limestone (C_m) rests on Woodside shale (TRw). Jointed rocks in foreground are siltstones of the Dinwoody formation.

FIGURE 3.—RED ROCK FAULT ZONE ALONG FRONT OF TENDOY RANGE, WEST OF U. S. HIGHWAY 91, BETWEEN LIMA AND DELL

Dark band at base of triangular facets is modern fault scarplet cutting through alluvial fans.

PLATE 5.—GEOMORPHIC FEATURES

FIGURE 1.—SUMMIT LEVEL EROSION SURFACE ACROSS BLACKTAIL RANGE

View looking southwest, up Jake Creek. Photo by E. Wm. Heinrich.

FIGURE 2.—LATE TERTIARY, GRAVEL-COVERED EROSION SURFACE IN TENDOY RANGE

View looking north. Folded Thaynes strata underlie horizontal Medicine Lodge limestone. Lima Peaks in background consist of Quadrant quartzite.

FIGURE 3.—GROUND, TERMINAL, AND RECESSIONAL MORAINES AND STREAM TERRACES IN BIG SHEEP CREEK BASIN

View looking north from Beaverhead Range, down Trail Creek and Nicholia Creek.

pian (Scott, 1935, p. 1031) and may be correlated with the greater part of the Brazer formation of the "Idaho facies."

Pennsylvanian System

Pennsylvanian rocks comprise the Amsden limestones and shales and overlying Quadrant quartzite. The Amsden is part of the "Montana facies" and is absent in the Beaverhead Range. In the Tendoy Range the thickness is estimated at more than 200 feet; 150 feet is present in the Snowcrest Range (Sloss and Moritz, personal communication), and in the Blacktail Range the formation is 90 feet thick. The Quadrant formation is exposed in each one of the mountain ranges (Pl. 5, fig. 2). About 1000 feet occurs in the eastern part of the Snowcrest Range (Condit, 1918, p. 111), in the Blacktail Range, and in the mountains north of Armstead. Sloss and Moritz (personal communication) measured more than 1700 feet in the western part of the Snowcrest Range. In the Tendoy Range more than 2600 feet was measured in Big Sheep Creek canyon. Thus, the Quadrant thickens rapidly to the southwest.

In some places a rather abrupt lithologic break separates the Amsden and Quadrant formations, but elsewhere the contact is gradational and appears conformable. Upward, the Quadrant grades into the Permian Phosphoria formation with no evidence of an unconformity. The local occurrence of a conglomerate with limestone pebbles in the basal unit of the Amsden suggests that a disconformity separates it from the underlying Big Snowy (?) group, but the contact is too poorly exposed to verify this.

No identifiable fossils were found in the Amsden formation. The formation is regarded as straddling the Mississippian-Pennsylvanian boundary (Sloss, 1946, p. 6). Thus, the lower Amsden, and possibly the entire formation, is correlative with the upper part of the Brazer formation in the southwestern corner of the region ("Idaho facies"). Except for some extremely scarce, poorly preserved fusulinids, no fossil remains occur in the Quadrant formation. The correlative Tensleep formation of Wyoming is dated as Lower Pennsylvanian (Scott, 1935, p. 1019), but the conformable relation between the Quadrant and Phosphoria formations indicates that the former also consists of Upper Pennsylvanian strata, a conclusion also reached by Sloss and Moritz (1951, p. 2165).

Permian System

The Permian System is represented by the Phosphoria formation. The formation has been divided into a chert member at the top underlain by an upper phosphatic member, an upper carbonate member, a lower phosphatic member, and a lower carbonate member (McKelvey, 1949, p. 274; Klepper, 1951, p. 62–66).

The westernmost outcrops of the Phosphoria formation in the Lima region are in the Tendoy Range, where it is well exposed along Big Sheep Creek (813 feet) and Little Sheep Creek (710 feet). Eastward the formation occurs in both the Blacktail and Snowcrest ranges (360 feet and 427 feet, respectively). The beds thin northward and eastward; 205 feet of Phosphoria is reported from Melrose, north of the Lima region (McKelvey, 1949, p. 273).

The Phosphoria conformably overlies the Pennsylvanian Quadrant formation. The contact with the overlying Dinwoody formation (Triassic) is too poorly exposed to permit study. Sloss and Moritz (1951, p. 2167; Moritz, 1951, p. 1784–1785) believe that sedimentation may have been continuous from Permian into Triassic time. Thus, the Phosphoria formation would represent the entire Permian system.

Triassic System

The sedimentary rocks constituting the Triassic system include the shales and siltstones of the Dinwoody and Woodside formations (Pl. 4, fig. 2) and the silty limestones of the Thaynes formation (Pl. 5, fig. 2), all Early Triassic. About 650 feet of Dinwoody is present along the West Fork of Blacktail Creek in the Snowcrest Range; along Little Water, Little Sheep (Middle Fork), and Deep creeks in the Tendoy Range the thickness is approximately 870 feet. Only 180 feet of Dinwoody strata occurs in the Blacktail Range. Thus, the formation thins eastward and northward. The Woodside formation is 122 feet thick in Little

Water Canyon and 104 feet south of Mt. Garfield, along Little Sheep Creek. Eastward, it thickens to 186 feet along Blacktail Creek in the Snowcrest Range, but pinches out to the north and is absent in the Blacktail Range. It is uncertain whether Woodside beds are present in the Armstead area. The thickness of the Thaynes formation south of Mt. Garfield is 681 feet, similar to other places in the Tendoy Range. In the Blacktail and Snowcrest ranges much of the formation has been removed by post-Thaynes erosion, but the thickness of the remaining strata, 955 and 636 feet, respectively, indicates eastward thickening.

Where present, the three Triassic formations are mutually conformable, but in the Blacktail Range, where the Thaynes rests directly on the Dinwoody formation, the contact is disconformable. A conformable relationship probably exists between the Dinwoody and the underlying Phosphoria formation. The Triassic strata are truncated by a regional unconformity. As a result, the Thaynes formation, overlain in the Lima region by the Jurassic Ellis group and Morrison formation, pinches out farther east where, in the absence of Woodside, the Ellis overlaps on the Dinwoody formation (Moritz, 1951, Fig. 15).

The occurrence of *Meekoceras* in rocks mapped as Dinwoody presents a problem. In Wyoming and Idaho strata containing an abundance of *Meekoceras* specimens have been regarded by Newell and Kummel (1942, p. 947; Kummel, 1943, p. 316–322) as the base of the Thaynes formation. The occurrence of these cephalopods, though not in quite such great numbers, in rocks of typical Dinwoody lithology in the Lima region illustrates the limitations of biostratigraphic zones in defining formations.

Jurassic System

Four Jurassic formations—the Sawtooth shale, Rierdon limestone and Swift sandstone of the Ellis group, and the overlying Morrison shale and siltstone—are recognized. None are present in the northwestern part of the region which forms part of a large area in southwestern Montana over which Jurassic strata are absent (Moritz, 1951, p. 1805; Schmitt, 1953, p. 367–369).

Like the rest of the Ellis group the Sawtooth is restricted to the central and southern portions of the Tendoy Range, where it is 150 to 200 feet thick. A section measured along the Middle Fork of Little Sheep Creek included 167 feet of Rierdon strata. In the Little Water Creek area the thickness is only 86 feet, and the Rierdon pinches out farther north and east. No natural outcrops of the Swift formation were found, but the formation is exposed in a trench dug by the U.S. Geological Survey in NE ¼ SW ¼ sec. 10, T. 13 S., R. 10 W., in Little Water Canyon. The Morrison formation occurs on the west slope of the Blacktail Range (130 feet) and on the southeast slope of the Snowcrest Range (155 feet). From there it thickens westward to 327 feet in the Tendoy Range south of Garfield Mountain. Northward in the Tendoy Range it thins rapidly; only 71 feet of Morrison is present in the Little Water Creek section, and the formation is absent in the northern part of the range.

The Ellis group overlies the Thaynes formation unconformably. The Sawtooth and Rierdon formations are mutually conformable. The Sawtooth is assigned to the upper Middle Jurassic (upper Bajocian and Bathonian), and the Rierdon belongs to the lowermost Upper Jurassic (lower Callovian) (Imlay *et al.*, 1948). The contact between the Rierdon and Swift formations represents a slight stratigraphic break as the Swift is regarded as Oxfordian, possibly in part late Callovian, according to the same writers. Imlay *et al.* assign the Morrison to the Kimmeridgian and the lower third of the Portlandian and consider the formation conformable to the Swift. An important hiatus is represented by the contact between the Morrison and the Triassic strata in the eastern part of the region, where the Ellis group is absent. The Morrison is disconformably overlain by the basal conglomerates of the Kootenai formation.

Cretaceous System

The Cretaceous System includes the Kootenai sandstone and Aspen shale. Cretaceous formations younger than Aspen may also be represented. Klepper (1951, p. 68–69) reports evidence that equivalents of the Wayan and Frontier formations and possibly also the

Mesa Verde may be present in the Ruby River valley, east of the Lima region. In the absence of such evidence in the Lima region all post-Kootenai beds were tentatively mapped as Aspen. In the Tendoy Range nearly 1500 feet of Kootenai strata is present in the area of Sawmill Creek, southeast of Lima. Farther northeast, a well drilled near Junction Creek encountered 1155 feet, and the formation thins to about 900 feet in the Snowcrest Range. In the southeastern part of the Tendoy Range more than 3500 feet of Aspen is present although pre-Tertiary erosion has removed the upper portion. Near Sawmill Creek a characteristic badland topography is developed on the bentonitic beds, which are weathered in vivid colors and earn these localities the name of "Paintpots." To the east the formation occurs on the south flank of the Snowcrest Range, but farther north it has been removed.

The contact between the Kootenai and Aspen is gradational, and the formations appear to be conformable. North of the Lima region the Kootenai overlaps pre-Jurassic rocks (Moritz, 1951, Fig. 15); hence, the base of the formation represents an erosional unconformity. The Kootenai formation, according to all authorities, belongs to the Lower Cretaceous, but the age of the Aspen is disputed. Brown (1933, p. 3) states that the Aspen is older than Mesa Verde and younger than Dakota and assigns it to the lower Upper Cretaceous. Cobban and Reeside (1951, p. 1892) placed the Mowry and the equivalent Aspen formations in the Lower Cretaceous, but Yen (1952, p. 762) has concluded that the Bear River formation, which underlies the Aspen in Wyoming, is Cenomanian (lower Upper Cretaceous). LaRocque and Edwards (1954, p. 325), on the basis of stratigraphic studies in Teton County, Wyoming, suggest the possibility of a complex intertonguing relationship between the Bear River and Aspen formations but refrain from presenting this as a definite solution to the age problem. Fossils from the Lima region (Table 1) are chronologically inconclusive, but ecologically significant; they indicate lacustrine and swampy environments during at least part of the Cretaceous.

Tertiary System

The Tertiary includes the Beaverhead conglomerate of probable Paleocene age and a variety of post-Laramide sediments collectively referred to as "basin beds." Thin gravel deposits, possibly Pliocene, rest on a high Tertiary erosion surface. The volcanic rocks associated with the "basin beds" are discussed under Igneous Rocks.

Lowell and Klepper (1953) gave the name Beaverhead formation to a thick, coarse, dominantly red conglomerate with lesser amounts of sandstone and local fresh-water limestone in Beaverhead County, Montana (Pl. 2, fig. 2). In the type locality near the mouth of McKnight Canyon, Tendoy Range, the calculated thickness is 9700 feet. Top and bottom of the formation are cut off by faults. The Beaverhead formation is widely distributed across the Lima region but is absent in the Beaverhead Range. The siltstones and sandstones in the east are difficult to distinguish from the underlying Aspen formation. No complete section of the Beaverhead was found; in most places the top has been removed by erosion.

In the absence of diagnostic fossils, dating of the Beaverhead formation is based on stratigraphic and structural data. A profound angular unconformity of considerable relief separates the Beaverhead from rocks as young as the Upper Cretaceous Aspen and possibly Mesa Verde formations. The conglomerates, evidently supplied by early Laramide uplifts, have been involved in folding and thrusting during advanced stages of the Laramide orogeny. Consequently, the formation must be latest Cretaceous or Paleocene, or both. Eardley (1951, p. 319) states that the Beaverhead seems tectonically similar to the Paleocene Pinyon conglomerate of Wyoming and tentatively considers it of the same age. Lowell and Klepper (1953, p. 240) believe the age is probably early Paleocene. A conglomerate apparently very similar to the Beaverhead has been described in the Madison Range of Montana by Peale (1896, p. 3), who tentatively assigned it to the earliest Tertiary. This formation, referred to as Sphinx conglomerate by

Peale, probably is a time correlative of the Beaverhead also.

Among the "basin beds" three units of formational rank have been established: the Eocene Sage Creek formation (Douglass, 1903, p. 145), the Oligocene Cook Ranch formation (Wood, 1934, p. 254), and the Miocene Blacktail Deer Creek formation (Douglass, 1902, p. 243). The writers have mapped two additional units, provisionally termed Muddy Creek beds (Oligocene ?) and Medicine Lodge beds (Miocene ?). The latter may prove to be sufficiently widespread to deserve formational rank, or future work may justify inclusion of the two units in the Cook Ranch and Blacktail Deer Creek formations, respectively.

The type locality of the Sage Creek formation is about 8 miles due north of Lima. The formation is distributed over an area of about 4 square miles between the valley of Sage Creek and the Red Rock Mountains to the north. The thickness of the Sage Creek beds is uncertain but is estimated to be at least 300–400 feet. At the type locality the exposed beds probably are not more than 100 feet thick. The Sage Creek formation rests with distinct angular unconformity on the Beaverhead formation. The basal basalts seem to lie in old stream valleys. The formation is overlain unconformably by the Cook Ranch formation, whose basal unit consists in part of reworked Sage Creek debris. Mammal remains collected from the Sage Creek (Table 1) indicate a late Eocene age for the formation (Wood, 1934, p. 255). The Sage Creek is the only known Eocene deposit in Montana.

The Cook Ranch formation was defined and described by Wood (1934, p. 254) from exposures in the valley of Sage Creek. The Cook Ranch sediments and associated volcanic materials are present along the crest and east side of the Red Rock Mountains, along the valley of Sage Creek and adjacent to, as well as in, the Blacktail Range. The thickness is variable. Wood (1934, p. 254) estimates that 125 feet of Cook Ranch strata is exposed at the type locality, but the formation thickens rapidly northward to an estimated 1000 feet or more. The Cook Ranch is unconformable with the overlying Miocene Blacktail Deer Creek formation and the underlying Eocene Sage Creek formation and older rocks. On the basis of fossil data (Table 1) Wood (1941, p. 18) assigned the beds to the middle Oligocene (Orellan stage).

The Muddy Creek beds are restricted to the intramontane Muddy Creek and Muddy Hole basins in the Tendoy Range. Their thickness is unknown, but a well drilled to a depth of 1000 feet in Muddy Creek Basin did not encounter "hard rock" (Winchester, 1923, p. 87). The Muddy Creek beds are unconformable with Paleozoic rocks and are overlain by andesites and basalts. The sediments were involved in post-Laramide faulting. They are tentatively correlated with the middle Oligocene Cook Ranch formation farther east, on the basis of similar volcanic materials abundantly associated with both units, and general similarity in appearance.

The Blacktail Deer Creek formation was defined by Douglass (1902, p. 243) for exposures of light-colored, fine-grained sediments on Blacktail Creek, about 30 miles southeast of Dillon, Montana. Hibbard and Keenmon (1950, p. 194–198) have described the type section in detail. The total thickness of the formation is estimated to be in excess of 2000 feet. At the type locality, where base and top are not exposed, the thickness is about 400 feet. The Blacktail Deer Creek formation lies with angular unconformity on strata as young as the middle Oligocene Cook Ranch formation. No Tertiary sediments overlie the Blacktail Deer Creek beds. The fossil assemblage (Table 1) dates the Blacktail Deer Creek formation as lower Miocene, although Douglass (1902, p. 239) had considered it Oligocene.

The Medicine Lodge beds occupy the intermontane basin between the Beaverhead and Tendoy ranges and a small plateau in the Tendoy Range (Pl. 5, fig. 2). Their maximum thickness is calculated to exceed 5000 feet. A partial section, measured in Big Sheep Creek basin, is more than 2600 feet thick. No diagnostic fossils have been found in the Medicine Lodge beds of the Lima region, but J. Leroy Kay (personal communication) states that lithologically and stratigraphically similar sediments near (North) Bannock Pass, farther north in the depression east of the Beaverhead Range, contain remains of *Merychippus*, a late

Miocene horse. The Medicine Lodge beds are underlain by basaltic and andesitic lavas similar to those overlying the Oligocene (?) Muddy Creek beds farther west. For these reasons the strata are tentatively assigned to the Miocene. They may be equivalent to the Blacktail Deer Creek formation in the eastern part of the Lima region.

The latest Tertiary, or possibly earliest Quaternary, deposit consists of a thin veneer of gravel, composed of well-rounded to subrounded pebbles and cobbles which rest on a high erosion surface in the Tendoy and Beaverhead ranges. They are discussed in the section on Geomorphology.

Quaternary System

The Quaternary System includes the late Pleistocene Bull Lake (?) and Pinedale (?) moraines (Pl. 5, fig. 3) and the Recent terrace gravels, fanglomerates, landslide deposits, and stream-bed alluvium.

The drift deposits are discussed under Geomorphology. Terrace gravels, consisting of well-sorted and well-rounded pebbles, are widely distributed across the intermontane basins. Large fanglomerates skirt the northeast flanks of the Tendoy and Blacktail ranges and are associated with late Cenozoic faulting. Landslides are found in several of the major canyons, notably those of Nicholia Creek, Kate Creek, Blacktail Creek (Middle and West forks), and Cottonwood Creek, and at the head of Clark Canyon. Silty and clayey alluvial deposits are distributed along the valley flats of the main streams.

Igneous Rocks

Introduction

Both intrusive and extrusive igneous rocks are present in the Lima region. A large granitic pluton of Laramide age is exposed in the Beaverhead Range along the Continental Divide and may be consanguineous with a small granitic stock at the southwestern end of the Snowcrest Range. Many tabular intrusive bodies, including quartz veins, ultramafic sheets, lamprophyre, and diabase dikes, and other acidic and basic hypabyssal bodies crop out. All but two in the Ruby and Blacktail ranges are of Tertiary or probable Tertiary age.

Extrusive rocks, both lavas and pyroclastics, cover large portions of the region. They range in age from Eocene through Pleistocene (?) and in composition from rhyolites to basalts. These rocks are confined chiefly to the topographic basins and the slopes of the adjacent mountains, but in a few places they are found near the crests of the mountains. Thus, the volcanic rocks comprise a wide range of petrographic types, stratigraphic levels, and geographic distribution. A strict classification of the various units is desirable (Table 3). Many of the Tertiary extrusive rocks are intimately associated with the basin sediments in the region, and have been assigned to stratigraphic units named after the corresponding sedimentary units. Thus, in the area east and north of U. S. Highway 91 it is possible to distinguish the Sage Creek basalts, Cook Ranch rhyolites, and associated volcanics, and Blacktail Deer Creek basalts. West of the highway are the Muddy Creek and Medicine Lodge volcanic rocks, in addition to Pliocene (?) rhyolites which have been named after Edie School. Other occurrences cannot be readily placed stratigraphically, and their assignment to one of the above units, mostly based on petrographic similarities, is questionable. Several occurrences of volcanic rocks are so intimately associated with the basin sediments that they were mapped with them. This is particularly true for the Muddy Creek pyroclastics and for part of the Cook Ranch volcanics.

The Pliocene to Pleistocene basalts of the Snake River region extend into the southeastern part of the Lima region. Another basaltic flow of probable late Cenozoic age is present east of Armstead in the Clark Canyon area.

Intrusive Rocks

Beaverhead granite.—The name Beverhead granite has been assigned to the plutonic complex which crops out in the Beaverhead

TABLE 3.—IGNEOUS ROCKS IN THE LIMA REGION

Epoch	Petrographic type	Names and locations west and south of U.S. 91	Names and locations east and north of U.S. 91	Regional correlation
Pleistocene	Basaltic lavas	Snake River lava in South Medicine Lodge Basin	Clark Canyon lava near Armstead	Snake River lavas of central Idaho
Pliocene	Rhyolitic lavas	Edie School lava in South Medicine Lodge Basin		Tertiary Late lavas of central Idaho
Miocene	Basaltic and andesitic lavas; minor volcanic agglomerate and tuff. Some basic hypabyssal bodies	Medicine Lodge volcanics on slope of Beaverhead Range, in Big Sheep Creek, North Medicine Lodge and Muddy Creek basins and south of Armstead	Blacktail Deer Creek basalts in and between Ruby, Blacktail and Snowcrest ranges	Challis volcanics and Tertiary Early lavas of central Idaho
Oligocene	Rhyolitic tuffs and agglomerate; lesser amounts of lava. Some acid hypabyssal bodies	Muddy Creek volcanics in Muddy Creek, Muddy Hole, Red Rock and Big Sheep Creek basins	Cook Ranch lavas and volcanics in area of Red Rock Mountains and Blacktail Range	
Eocene	Basaltic lavas		Sage Creek lava on south end of Red Rock Mountains	
	Granitic to quartz monzonitic plutons. Lamprophyre dikes, quartz veins and aplites	Beaverhead granite batholith in Beaverhead Range	Snowcrest granite stock at southwest end of Snowcrest Range	Various intrusives in central Idaho
Paleocene				
Precambrian	Acidic to ultra-basic hypabyssal bodies	Post-tectonic pegmatites, aplites, and quartz veins associated with Dillon granite gneiss	Ultra-mafic sheet in Blacktail Range Diabase dike in Ruby Range	

Range along the Idaho State line. The body is elongate parallel to the trend of the mountain range and is more than 6 miles long and 2 miles wide in outcrop. It is cut by coarse quartz veins and by small aplite dikes of the same composition as the main body, but of much finer texture. These rocks are more resistant than the surrounding Paleozoic limestones and support the highest peaks along the Continental Divide. A coarse talus covers the high slopes.

The bulk of this igneous complex is a moderately coarse-grained granite ranging from light gray to pink and light green. Only quartz

and pinkish feldspar can be recognized megascopically.

Under the microscope specimens from the southern portion of the intrusive reveal the presence of quartz and K-feldspar in about equal amounts. Zircon is the only significant accessory mineral. Some hematite and clusters of magnetite were observed. The feldspar is largely altered to a dense aggregate of very small sericite flakes. Farther north, at the head of Meadow Creek, a variety of granite contains abundant microcline.

At the northern limit of the outcrop oligoclase constitutes more than one-third of the total feldspar, and the rock approaches a quartz monzonite. Microcline is abundant, and quartz makes up about 40 per cent of this rock. Micrographic intergrowths are abundant, and the microcline occurs in perthitic intergrowths with plagioclase. The dark-mineral content is much higher than farther south, and many flakes of biotite can be observed, most of which are greatly altered to chlorite and magnetite. Most of the feldspar is strongly sericitized and kaolinized. Zircon is the most abundant accessory.

The Beaverhead pluton is the same as that described by Umpleby (1913a, p. 42) and Shenon (1928, p. 9), who mapped it on the Idaho side of the State line. A small intrusive of the quartz monzonite variety was observed by Anderson and Wagner (1944, p. 9) a few miles to the south, in Lemhi County, Idaho. None of these intrusives seem to be related to either the Idaho or the Boulder batholith; in general they are more alkalic. The Beaverhead pluton has an intrusive relationship to Laramide folds believed to have originated in early Eocene time. It also cuts through a high-angle fault of probable Laramide age. Volcanic rocks, thought to be Oligocene to Miocene, overlie the granite and were extruded on its eroded surface, as can be observed at the head of Meadow Creek. Some volcanic agglomerates contain boulders derived from the granite. This evidence points to a probable Eocene age for the pluton.

Snowcrest granite.—The name Snowcrest granite is applied to a small stock which crops out at the southwestern end of the Snowcrest Range, in section 20, T.13 S., R.7 W., Montana. The granite is strongly jointed and disintegrates into small, angular fragments. In hand specimen abundant quartz and feldspar, in addition to a little biotite, can be observed. The rock is coarse-grained and in thin section consists of about 54 per cent orthoclase, 19 per cent sodic plagioclase, 25 per cent quartz and 2 per cent biotite. The feldspar is strongly kaolinized. The rock occupies a position somewhat intermediate between the two varieties of the Beaverhead granite discussed above, except that there is less quartz in the Snowcrest.

The Snowcrest intrusive cuts through Mississippian rocks and is unconformably overlain by the lower Miocene Blacktail Deer Creek formation, which contains fragments of granite in a basal breccia. The Snowcrest granite may be consanguineous with the Beaverhead pluton and may be tentatively considered Eocene.

Hypabyssal bodies.—Quartz veins are the most common type of hypabyssal body. Several such veins occur in the pre-Beltian rocks of the Beaverhead and Tendoy ranges. One cuts through the plane of the Laramide Cabin thrust into underlying Mississippian rocks. In the Armstead area quartz veins are intrusive into rocks as young as Triassic. Evidently, some of the veins are late Laramide or post-Laramide and may be related to the Beaverhead granite. Other veins may be older. The veins are largely milky white and bluish quartz. Microscopic examination reveals the presence of chalcedony, serpentine, chlorite, sericite, calcite, and iron ore minerals in varying quantities.

Dikes of dark, basic rocks, all less than 30 feet wide and less than 100 feet long, are found in the Beaverhead Range. Their extent is exaggerated on the geologic map (Pl. 1). All are intrusive into Mississippian rocks. The main primary mineral is a titaniferous augite, largely altered to an aggregate of chlorite, carbonates, iron oxides, and quartz grains. Feldspar is present, but most of it has been altered to sericite. Other primary minerals are biotite, magnetite, apatite, and, locally, sphene. The alteration products constitute an important part of the rock, and the original composition is consequently in doubt. The

rocks may be lamprophyric varieties. The age of the dikes is also uncertain, except that they are post-Mississippian. They occur in proximity to the Beaverhead pluton, but evidence for a genetic relationship is lacking.

Another hypabyssal body supports a cliff at the northern end of Muddy Creek Basin, along Sourdough Creek. In a hand specimen small phenocrysts of feldspar can be seen in a light pinkish groundmass. Microscopic examination shows that the feldspar is a fresh oligoclase. A few long flakes of biotite are conspicuous. The groundmass is very fine-grained and consists mostly of kaolinized orthoclase, quartz, and some small grains of hornblende. The rock has the approximate composition of rhyolite and because of its porphyritic texture is considered to represent the intrusive equivalent of the rhyolitic tuffs of Muddy Creek Basin.

An andesitic rock mass, intrusive into Triassic rocks, occurs in Garfield Canyon, southwest of Armstead. It has the same composition as the prevailing Tertiary lava flows in this vicinity, but a coarser texture. This intrusive is believed to be a feeder to the flows. In section 13, T. 14 S., R. 11 W., Montana, a porphyritic rock occurs in the pre-Beltian. The field evidence is inconclusive, but the texture suggests that this may be another hypabyssal body. The rock is grayish green and contains phenocrysts up to 4 mm long, consisting of oligoclase (?) largely replaced by sericite; some crystals are preserved only in outline. The same feldspar occurs in small laths in the groundmass, the remainder of which includes many small magnetite grains, chlorite, biotite, colorless mica, and a little hematite. No quartz was observed. The rock is so altered that no definite identification is possible, but it is believed to be andesitic. It may be related to the Tertiary volcanics in the area.

A horizontal ultramafic sheet, approximately a mile long and 50 to 200 feet wide, is exposed along Jake Creek in the Blacktail Range. Sinkler (1942, p. 137) mentions this occurrence and describes in detail similar bodies in the Ruby Range several miles to the northeast, which are probably consanguineous. These intrusives consist of nickeliferous peridotite, much of it rich in enstatite (saxonite or harzburgite, and enstatolite), and partly altered to serpentine. Locally encrustations of annabergite occur (Sinkler, 1942, p. 138–141). The rock is said to contain up to 2 per cent nickel and a trace of platinum. It is surrounded by rocks of the pre-Beltian Pony group. Lack of metamorphism indicates a post-Cherry Creek age. The sheet may represent a late Precambrian intrusion, but the field evidence is inconclusive.

A diabase dike occurs in the Ruby Range, in the northeastern corner of the Lima region. It cuts the Pony gneisses, is unmetamorphosed, and is tentatively considered late Precambrian.

Extrusive Rocks

Sage Creek basalts.—Several small basalt flows occur at the base of the Sage Creek formation (upper Eocene), in T. 12 S., R. 8 W., Montana, north of Sage Creek. Similar flows crop out 2 miles to the northwest, where they are overlain by sediments of the middle Oligocene Cook Ranch formation. Sage Creek sediments are absent in this vicinity. The same condition exists farther north, on the southwest side of the Blacktail Range, sections 2 and 11, T.10 S., R.9 W., Montana. Both occurrences are tentatively considered erosional remnants of the Sage Creek basalt. The rocks are all badly weathered.

Cook Ranch rhyolites.—Pink and buff rhyolitic lava flows are exposed on the southwest side of the Blacktail Range. They are intercalated between pyroclastic deposits that have been correlated with the middle Oligocene Cook Ranch formation. One thin section discloses a glassy groundmass with unoriented phenocrysts, including sanidine, albite, corroded biotite, and quartz. The quartz is locally embayed. Accessory minerals include magnetite and apatite. Limonite is an alteration product.

Other rhyolites crop out over a large area at the southeast end of the Blacktail Range, in the vicinity of Prices Canyon. They are interbedded with rhyolitic tuffs and agglomerates which were not mapped separately in this area. Most of the lavas are red to brown and

largely aphanitic. Most of the flows are resistant to erosion, whereas the tuffaceous beds are resistant only where silicified. In thin section one specimen of red rhyolite shows abundant volcanic glass with microlites arranged in a poorly defined flow structure. Quartz and andesine form small phenocrysts. Accessory minerals include apatite, zircon, magnetite in small, evenly scattered grains, and green biotite. Hematite and limonite are alteration products. These extrusives overlap on the Precambrian core of the Blacktail Range and are overlain by sediments of the lower Miocene Blacktail Deer Creek formation. Correlation of these flows with those in the Cook Ranch formation is likely. The Cook Ranch rhyolites probably include a number of different flows, and petrographic variations as described above are to be expected.

Cook Ranch volcanics.—The term Cook Ranch volcanics designates beds of volcanic tuff and white, acidic volcanic breccia associated with sediments of the Cook Ranch formation (middle Oligocene). Some volcanic ash is present in the Cook Ranch of the type locality, at the south end of the Red Rock Mountains. Northward, the ash content increases; and the formation grades into interbedded tuffs, volcanic breccias, and agglomerates at the foot, on the slopes, and near the crest of the Blacktail Range. A disconnected patch of rhyolitic pyroclastics with minor lava occurs at the head of Cottonwood Creek, in and around section 29, T.10 S., R.7 W. The over-all distribution of the Cook Ranch volcanics suggests that these rocks once covered the entire southeastern end of the Blacktail Range.

Blacktail Deer Creek basalts.—Several basalt flows are associated with the Blacktail Deer Creek formation (lower Miocene). They are exposed in a northeast-southwest zone northwest of the Snowcrest Range, between U. S. Highway 91 and Blacktail Creek. Hand specimens from several localities are composed of a dark, locally vesicular groundmass with phenocrysts of feldspar. Microscopically the feldspar consists of an acid labradorite and occurs in the groundmass as well. In two of the four thin sections examined augite forms small phenocrysts. Field relations as well as petrographic similarity indicate that these flows all belong to the same phase of volcanism.

Similar basalts are exposed over a large area along Clark Divide, about 8 miles east of Armstead. They rest unconformably on Cook Ranch deposits and are beveled by an erosion surface believed to be late Tertiary. Small patches of basalt overlie Precambrian rocks and Cook Ranch volcanics in the southeastern part of the Blacktail Range, at the head of Cottonwood Creek. These lavas contain acid labradorite, augite, some olivine, and lesser amounts of magnetite. Chlorite has locally replaced plagioclase. There is some petrographic similarity to the Blacktail Deer Creek basalts. The same is true for basalt flows in the Ruby Divide area, in the northeastern corner of the Lima region, which rest on Blacktail Deer Creek sediments. A tentative placing of the rocks in these three localities in and near the Blacktail Range within the Blacktail Deer Creek basalt sequence seems justified. If these correlations are correct, the lower Miocene basalts must have once extended widely across portions of the Blacktail and Ruby ranges and adjacent areas.

Finally, remnants of lava scattered on the south slope of the Snowcrest Range contain small phenocrysts of augite and of a basic andesine, embedded in a matrix consisting partly of glass and partly of plagioclase microlites. A little quartz is present but may be secondary. Accessory minerals are magnetite and apatite. The lavas rest on various Cretaceous sediments. They have tentatively been included with the Blacktail Deer Creek lavas, although they appear to be andesites or quartz andesites rather than basalts. The correlation is questionable.

Muddy Creek volcanics.—The term Muddy Creek volcanics is used for acidic tuffs interbedded with Tertiary sediments in Muddy Creek and Muddy Hole basins. The tuffs range from white and light gray to pink and buff. Fresh-water gastropods, bedding, and lamination indicate their water-laid origin. Thin sections reveal that the relative abundance of glass, individual crystals, and rock fragments varies considerably from place to place. All

tuffs are rhyolitic with quartz, sanidine, oligoclase, and biotite as most important minerals, all embedded in brown glass. Accessory minerals are zircon, apatite, chlorite, calcite, allanite (?), and magnetite.

The Muddy Creek sediments associated with the tuffs are believed to be middle Oligocene. They are overlain by basalts and andesites similar to those in North Medicine Lodge and Big Sheep Creek basins farther west, which are considered late Oligocene or early Miocene. Near Island Butte, in the northern part of Big Sheep Creek Basin, a rhyolitic lava flow underlies these basalts and andesites and is therefore correlated with the Muddy Creek volcanics. This rock is greenish gray and dense. Phenocrysts are few and small, including mostly sanidine with some plagioclase. They are embedded in a cryptocrystalline groundmass, presumably consisting largely of quartz. Secondary quartz and calcite are present.

Another rhyolitic lava occurs in Red Rock Basin, south of Kidd. This rock is dense and is light gray, brown, and pink. Phenocrysts of sanidine and quartz are embedded in a groundmass of irregular grains of quartz, fewer feldspar laths, and a large quantity of brown volcanic glass. Flow structures are developed in several places. Locally, volcanic ashes are admixed, lending a tuffaceous appearance to the rock. Correlation of this rhyolite with the Muddy Creek volcanics is tentative and is based only on the relative proximity of the two. The flow is surrounded by Quaternary deposits. It may represent a link between the Muddy Creek and Cook Ranch volcanics, which are thought to be time correlatives on the basis of petrographic similarity and similar stratigraphic position.

Medicine Lodge volcanics.—The term Medicine Lodge volcanics is applied to intermediate and basic lavas and tuffs overlying the sediments and rhyolitic tuffs of Muddy Creek Basin and underlying the Medicine Lodge sediments in Big Sheep Creek Basin. Microscopic examination reveals both basalts and andesites. The two cannot be separated in the field. Microscopically, they are distinguished by a slightly different plagioclase composition in the groundmass.

The most common rock type in this series is a dense and slabby, dark-gray to greenish-black lava which weathers reddish brown. Dark-green phenocrysts of pyroxene are embedded in the aphanitic matrix. These rocks are dominant on the northeastern slopes of the Beaverhead Range. Many local varieties of this general type occur in the vicinity. One other type of basaltic lava is sufficiently common to deserve separate description. This rock is bluish-green and porphyritic. Greenish-black hornblende phenocrysts, up to half an inch long, are oriented in parallel planes, lending a platy habit to the rock. This rock is exposed north of Meadow Creek and in the vicinity of Island Butte, and a bright-green, slightly vesicular variety of the same type is found east of Nicholia Creek, near Henderson Gulch. Basaltic and andesitic lavas of these general types also occur along North Medicine Lodge Creek, on the north, west, and east sides of Muddy Creek Basin, and in the area south of Armstead. They are tentatively correlated with the Medicine Lodge volcanics. All flows in this group are relatively resistant to erosion. They are strongly jointed and break up in large blocks. In places, flows are preserved as tongue-shaped geomorphic features.

Microscopic examination of the common, dark-greenish variety reveals that the plagioclase microlites in the groundmass are between andesite and labradorite. They make up more than 80 per cent of the microcrystalline matrix. The remainder of the groundmass consists largely of magnetite in the form of fine dust and small crystals and minor amounts of tiny augite grains. Phenocrysts, which make up approximately 30 per cent of the rock, consist dominantly of pale-violet idiomorphic crystals of augite. Labradorite occurs in scattered lath-shaped crystals up to 1 cm. long. Accessory minerals include apatite, magnetite, and, locally, hypersthene and hornblende. Basaltic varieties contain some olivine. Many feldspar crystals show strong effects of magmatic resorption by rounded corners and matrix-filled embayments. The mineral is slightly chloritized with liberation of a little calcite. Serpentine replaces augite and olivine, liberating small amounts of magnetite, hematite, and quartz. The rock is characteristically glomeroporphyritic due to a tendency of phenocrysts to occur in clusters. Indistinct trachytic texture can be observed in places.

The bluish-green, porphyritic rock contains phenocrysts of brown basaltic hornblende. It is

strongly corroded, and many grains are rimmed by alteration products, including chlorite, biotite, calcite, magnetite, and hematite. Labradorite and brown hornblende are dominant in the microcrystalline matrix, which constitutes about 80 per cent of the rock. Augite and magnetite are locally abundant, the former partly altered to chlorite. The bright-green variety of this rock is intensely chloritized and contains more feldspar phenocrysts and abundant apatite.

The Medicine Lodge volcanics also include basaltic pyroclastic materials. Basalt tuffs are found in Big Sheep Creek Basin along the east side of Nicholia Creek north of the junction with Trail Creek, and in the Meadow Creek area. They are intercalated in the lavas and range from light purple to yellow, light gray, and white. Those along Nicholia Creek are lithic tuffs consisting largely of microscopic fragments of microcrystalline basaltic lava and some quartzitic sedimentary material, held together by brown volcanic glass. The pyroclastics along Meadow Creek are dominantly crystal tuffs with separate grains of hornblende, labradorite, and, locally, biotite, augite, and magnetite; volcanic glass occurs in places. Subangular lapilli, up to 2 inches long, are embedded in the ash. Finally, basaltic volcanic agglomerates occur in several places in association with Medicine Lodge extrusives. Four exposures were found near Nicholia Creek, on the northeast slope of the Beaverhead Range. Some are surrounded by lava flows, but others are isolated. One exposure occurs in the Tendoy Range, east of Muddy Creek Basin. The particles are gray, brown, and red, subrounded to subangular, and range from pebbles to boulders as much as 6 feet in diameter. In thin section the matrix is similar to the basaltic tuffs. Most of the pebbles and boulders consist of basic lava, but pieces of Mississippian limestone and of quartzite and granite are also found.

The age of this series of basaltic and andesitic rocks is uncertain. The lavas on the slope of the Beaverhead Range disappear northward below the Tertiary sediments of Medicine Lodge Basin, which probably are Miocene. The basalts and andesites near Muddy Creek overlie tuffs and sediments tentatively assigned to the middle Oligocene. If these assignments are correct, the Medicine Lodge volcanics are late Oligocene to early Miocene. Other considerations support this. Anderson and Wagner (1944, p. 11) describe basalts similar to those near Nicholia Creek from the southwest slope of the Beaverhead Range. They correlate these lavas with the "Tertiary Early Lavas" described by Kirkham (1927, p. 31–33) and with the Challis volcanics studied by Ross (1934, p. 46–53; 1937, p. 49–68; 1947, p. 1120) in central Idaho. Ross has assigned an Oligocene to early Miocene age to the Challis volcanics. Conceivably the Medicine Lodge volcanics form a link between the Challis volcanics and "Tertiary Early Lavas" of central Idaho and the early Miocene Blacktail Deer Creek basalts in the eastern part of the Lima region. This would indicate extremely widespread eruptive activity of basic and intermediate character during this period.

Edie School rhyolites.—Rhyolitic lavas overlie the mid-Tertiary sediments of South Medicine Lodge Basin and are also found on a small plateau in the upthrown fault block of the Tendoy Range, near Bannock Pass. These rhyolites are overlapped by the Snake River basalts of late Pliocene to Pleistocene age in the southeast corner of the Lima region. They are, therefore, believed to be Pliocene. Stratigraphically and petrographically they correspond to similar rhyolites bordering the Snake River plains of central Idaho (Kirkham, 1927, p. 33–38; Stearns *et al.*, 1938, p. 33–41; 1939, p. 20–35) which Kirkham termed "Tertiary Late Lavas."

The Edie School lavas are slightly porous, pale-lavender, rhyolitic rocks consisting of an aphanitic groundmass with glassy-looking phenocrysts. The rocks weather in shades of brown. They attain their greatest thickness, estimated at 200 to 250 feet, in the southeast corner of the area and wedge out where they overlap on the flanks of the adjoining ranges.

Microscopic examination shows that a micro- to cryptocrystalline groundmass constitutes about 90 to 95 per cent of the total rock mass. The matrix consists mostly of orthoclase and quartz, in addition to some albite and tiny grains of hematite and magnetite. Limonite causes a brown stain. Phenocrysts consist of approximately 45 per cent orthoclase, 25 per cent albite, and 25 per cent quartz.

Many quartz crystals are euhedral. Locally quartz occurs in myrmekitic intergrowth with albite. The rest of the phenocrysts are scattered grains of augite and magnetite with some biotite, apatite and zircon. Alteration is slight, except for minor chloritization of augite and limonitization of hematite and magnetite. The rocks possess a distinctly trachytic texture.

According to Kirkham (1927, p. 37–38) the "Tertiary Late Lavas" belong in large part to the great sheets extending westward from Yellowstone Park. Some of the material was erupted from small acidic lava and tuff cones found in various localities in central Idaho. One old cone, north of the settlement of Small, about 10 miles south of the Lima region, may have been the source of the rhyolites in South Medicine Lodge Basin. Other cones may be concealed by the Snake River basalts. The lavas may represent welded tuffs of the "*nuée ardente*" type, though Hausen (1954, p. 11–12) discounts this mode of origin for many rhyolites in Idaho.

Snake River basalts.—The Edie School rhyolites of South Medicine Lodge Basin are overlapped by dark, slabby basaltic rocks with a vesicular, aphanitic matrix. Phenocrysts of feldspar, olivine, and augite range up to 1 cm in length. These basalts thicken to an estimated maximum of about 120 feet in the southeast corner of the Lima region. Still farther south they continue to thicken.

The feldspar is a basic, zoned labradorite and constitutes about 65 per cent of the phenocrysts; the remainder are olivivine (30 per cent) and augite (5 per cent). Labradorite and augite are comparatively fresh, but the olivine crystals have rounded, corroded corners and are partly altered to antigorite, which occurs in tiny fibers perpendicular to margins and fractures. Much magnetite, and some hematite and iddingsite was liberated in the process. A few grains of ilmenite were found. About 70 per cent of the holocrystalline matrix is small labradorite laths, and the remaining 30 per cent is mostly tiny grains of augite. No flow structures were observed.

On the northeast side of the Tendoy Range, between Lima and Monida, disconnected patches of basaltic lava cap small buttes. Petrographically, these lavas are similar to the basalts of South Medicine Lodge Basin and to those near Humphrey, on the southeast side of Monida Pass, both of which grade laterally into the basalts of the Snake River region in central Idaho. There is considerable dissimilarity between these rocks and other basalts in the Lima region. The Snake River basalts are considered late Pliocene to Pleistocene. This is also the apparent age of the basalts between Lima and Monida, as they overlie a surface of probable late Tertiary age. Hence, lateral gradation, as well as petrographic and stratigraphic evidence, justifies assigning the term Snake River lavas to these rocks. Usage of this term does not imply that any of the basalts in the Lima region are actual extensions of individual flows in the Snake River plains.

Most of the Snake River flows are believed to be local and to have issued from central volcanic sources (Kirkham, 1927, p. 38; Stearns *et al.*, 1938, p. 56–58; 1939, p. 24–36). It is possible that early flows, now concealed, extruded from fissures and are more widespread, as suggested by Kirkham.

Clark Canyon lavas.—A sheet of dense, aphanitic basaltic lava crops out along Clark Canyon, near Armstead. It occurs at the bottom of a well-defined modern canyon cut into rocks as young as the Oligocene Cook Ranch volcanics. This suggests that they are late Tertiary or Pleistocene, hence the approximate equivalent of the Snake River basalts.

STRUCTURE

Pre-Beltian Structures

Metamorphic structures, such as alignment of minerals, shearing zones, schistosity, and gneissic banding in the rocks of the Pony and Cherry Creek series and the Dillon granite gneiss, testify to considerable pre-Beltian deformation attended by high pressures and temperatures. In most of the Tendoy Range the foliation strikes approximately north and dips steeply west or is vertical. The Cherry Creek rocks at the headwaters of Warm Springs Creek, northwest of Muddy Creek Basin, dip uniformly 15° NE. In the area of Blacktail Creek the foliation in the Pony rocks strikes northeast, and plunging folds are in evidence.

Laramide Structures

General remarks.—The most prominent Laramide structural elements consist of northeast-trending folds, northwest- to north-trending folds, northwest-trending low-angle thrusts, and northwest-trending high-angle thrusts (Figs. 3–6; Pls. 1 and 6). The thrust planes all dip southwest. These structures have been produced in the above sequence as a result of progressive stages of deformation during the Laramide orogeny. A number of less prominent folds and high-angle normal and reverse faults are present which are believed to be Laramide, but whose exact genetic relationships to the other structures is not clear. Just beyond the eastern edge of the Lima region a northeast-striking thrust dips northwest and appears to be genetically associated with the northeast folds in the region. Northeast structural trends dominate east of U.S. Highway 91, whereas northwest trends are more prominent west of this line. Where both trends are mutually superposed the structural pattern is extremely complicated and, in places, open to more than one interpretation.

Northeast folds.—A major northeast trending anticline in the northeast is evident from the areal distribution of the Paleozoic and Mesozoic strata in the Blacktail and Snowcrest ranges. The pre-Beltian core of this giant fold, here referred to as Blacktail-Snowcrest anticline, has been stripped of pre-Laramide sediments in the southeastern part of the Blacktail Range as well as in most of the Ruby Range to the north. It is now partly concealed by overlapping Tertiary sediments. Consequently, the exact position of the anticlinal axis is uncertain. The youngest beds involved in the folding belong to the Upper Cretaceous Aspen formation. They are overlain in angular unconformity by the Beaverhead conglomerates, probably Paleocene. The crest of the anticline appears to be broad and flat. Dips in the northwest limb of the structure are about 15°, but the opposite flank is steeper, with dips ranging from 45° to 60°. Parallel synclines interrupt the dips of both limbs of the major fold.

The southeast flank of the Blacktail-Snowcrest anticline steepens northeasterly, and 3 miles beyond the Lima region the strata are vertical to overturned and are overridden from the northwest by the pre-Beltian core along a northeast-trending thrust (Klepper, 1951, p. 74–75, Pl. 16). Westward, the beds in this limb disappear under Cenozoic sediments in the vicinity of the Red Rock River. Beyond this point the northeast trend is not only obscured by the Tertiary cover, but is also complicated by the superposition of the younger, northwest-trending Lima anticline and Tendoy thrust fault. The belt of Paleozoic and Mesozoic strata reappears at the surface on the south, or upthrown, side of this thrust. The general trend of the beds is still northeast, except at the plane of thrusting where they bend sharply to the northwest and are overturned, dipping about 60° SW. (Pl. 6, cross section BB′). This is believed to be the result of drag along the thrust fault.

The southwest-trending beds in the Tendoy thrust sheet represent the southeast flank of the Garfield anticline, a partial continuation of the Blacktail-Snowcrest anticline. The northeast pattern of this anticline has been modified by later northwest movements. The beds in the east limb dip steeply and locally stand on end. Abrupt changes in strike have been produced by dislocation along a number of high-angle faults. The opposite flank of the Garfield anticline can be recognized along the West Fork of Little Sheep Creek, where the strata strike north. Apparently, the anticlinal axis approximately parallels the creek 2–3 miles to the east and pitches to the southwest. Like the Blacktail-Snowcrest anticline, the folded beds of the Garfield anticline are unconformably overlain by Beaverhead conglomerates.

Another northeast structure, the sharply folded Little Water syncline exists in the vicinity of Hidden Pasture and Little Water creeks. Its axis pitches steeply southwest. A northwest syncline has been superposed upon this structure, creating a heart-shaped structural basin. Two minor high-angle faults, with their upthrown sides on the northwest, further complicate the structure. The northern limb of the Little Water syncline is the steeper. Both limbs are truncated by block faults on the southwest and northeast sides, and their

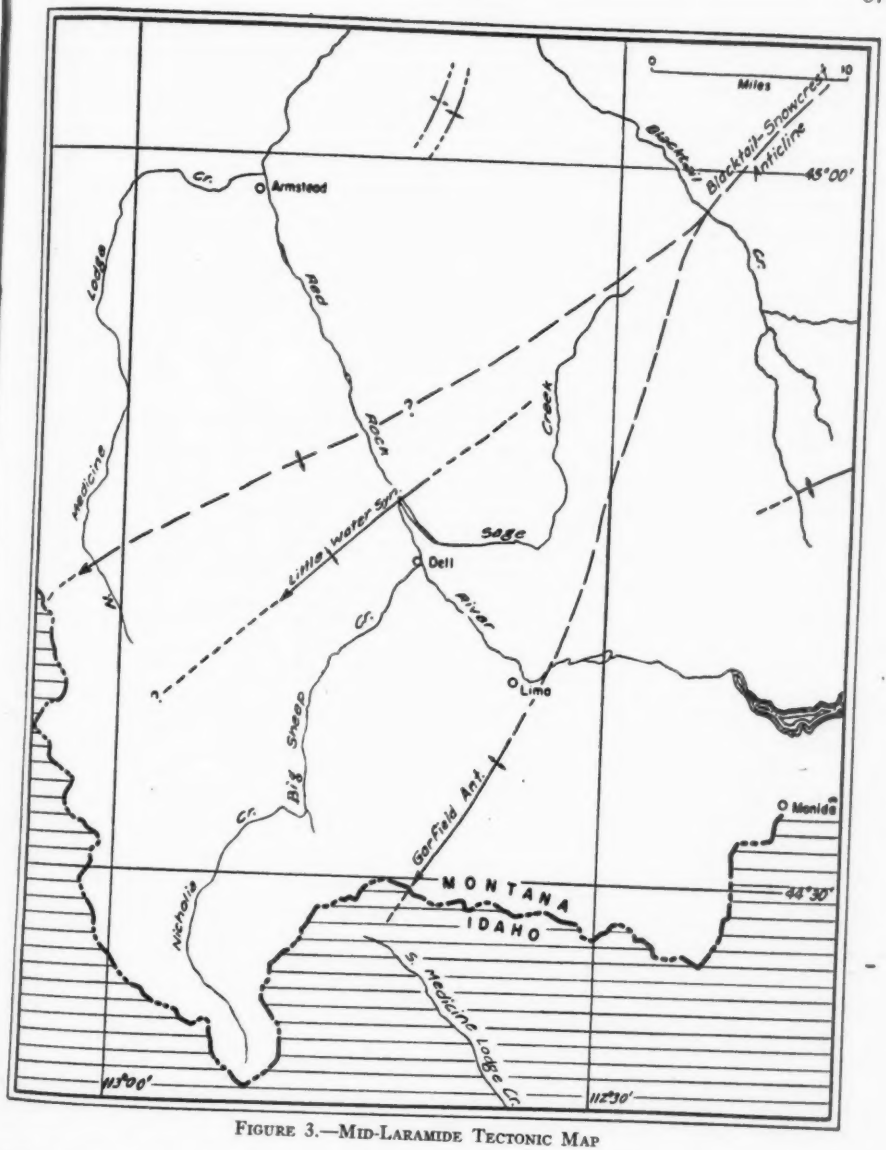

FIGURE 3.—MID-LARAMIDE TECTONIC MAP

continuations on the downthrown sides of these faults are covered by Cenozoic sediments.

A northeast-trending anticline adjacent to the Little Water syncline is indicated in the area south and southwest of Armstead by the reappearance of the Paleozoic and Mesozoic strata from under the Limekiln thrust. The beds strike northeast near the thrust outcrop. Farther away they swing to the north where they become involved in younger north- and northwest-trending structures. The structural pattern is complicated by these later trends, as well as by numerous high-angle faults. The axis of this northeast anticline is somewhere

between Little Water Creek and McKenzie Canyon, probably closer to the former, as the anticline is asymmetric to the southeast. The eastward extrapolation of this axis appears to link up with the northeast extension of the Garfield anticlinal axis in the vicinity of Sage Creek to form the main axis of the Blacktail-Snowcrest anticline, with the sharper and more steeply rising Little Water syncline dying out in this direction (Fig. 3).

Northwest folds.—Northeast-southwest compression severely complicated, and in part obliterated, the structural pattern set up by the earlier northwest-southeast forces. The later date of the northwest-trending structures is demonstrated by the fact that, in contrast to the northeast structures, they involve the Paleocene (?) Beaverhead conglomerates. The sharp turns of the beds in the Little Sheep Creek area from north and northeast to northwest were probably not present in the original Garfield anticline, but were produced by later compression athwart the early trend. Both structural directions can be recognized in a number of subsidiary, sharply curving folds in the Triassic beds along the West Fork of Little Sheep Creek. Similarly, a northwest syncline has been superposed on the northeast-trending Little Water syncline. Farther north, the beds that emerge with a northeasterly strike from beneath the Limekiln thrust make a sharp turn to form a northwest syncline.

West of Armstead these younger structures gradually curve north, and near the northern limit of the Lima region the strata are folded into a series of parallel synclines and anticlines with almost due northerly strike. The most prominent of these folds is the south-plunging Armstead anticline, in whose core are exposed the pre-Beltian Dillon granite gneiss and Pony group. Its steep east flank dips about 70°. Cross folds and transverse and longitudinal high-angle faults greatly complicate the structural pattern in this vicinity. Separated from this fold complex by the Tendoy thrust, but probably genetically related to it, is the parallel Clark Canyon syncline in the Beaverhead conglomerate east of Armstead, which connects the Armstead structural complex with the northwest flank of the Blacktail-Snowcrest anticline.

Northwest cross folds superposed on the Blacktail-Snowcrest anticline are indicated by the attitudes of the Paleozoic strata in the Blacktail Range near Ashbough Canyon and farther south. Two similar cross folds, the Red Rock syncline and Clover Creek anticline, exist in the southeast flank of this major structure, continuing into the Beaverhead conglomerates which truncate the flank. These relations confirm the later date of the northwest structures. The fold axes plunge southeastward.

Another northwest structure, the Lima anticline, occurs farther south. At the surface it involves only Aspen and Beaverhead rocks, which dip from 30° to 60°. The central part of the structure is mostly concealed by Quaternary deposits. Like the Red Rock syncline and Clover Creek anticline, the Lima structure was superposed on the southeast flank of the Blacktail-Snowcrest anticline, which, however, is concealed beneath the Beaverhead rocks in this vicinity. The outcrop pattern of the Beaverhead-Aspen contact, which had not been folded previously, indicates a northwest plunge for the Lima anticline. It is uncertain whether this plunge offsets the pre-existing southeasterly dip of the pre-Beaverhead strata in the Blacktail-Snowcrest anticline. It is possible that in the subsurface the Lima anticline pitches continuously southeastward.

Minor parallel folds occur on both flanks of the Lima anticline and farther south, in the upper Aspen and Beaverhead formations near the State line. In the area of Four Eyes Canyon, between Big and Little Sheep creeks, a northwest anticline has brought Devonian and Ordovician (?) rocks to the surface. Erosion has removed most of the central and northern portions of this structure, revealing the Beaverhead conglomerates across which it was thrust. The southwest limb dips steeply, in places 80°. The anticline pitches northwest, and it is probable, though not clear from the outcrop pattern, that there is also a southeast pitch and that the structure is domal.

Northwest folds occur within the Carboniferous strata in the southwest part of the region, such as north of Four Eyes Canyon, west of Muddy Creek and Muddy Hole basins, and west of Big Sheep Creek Basin. In places the beds are intensely contorted, probably in

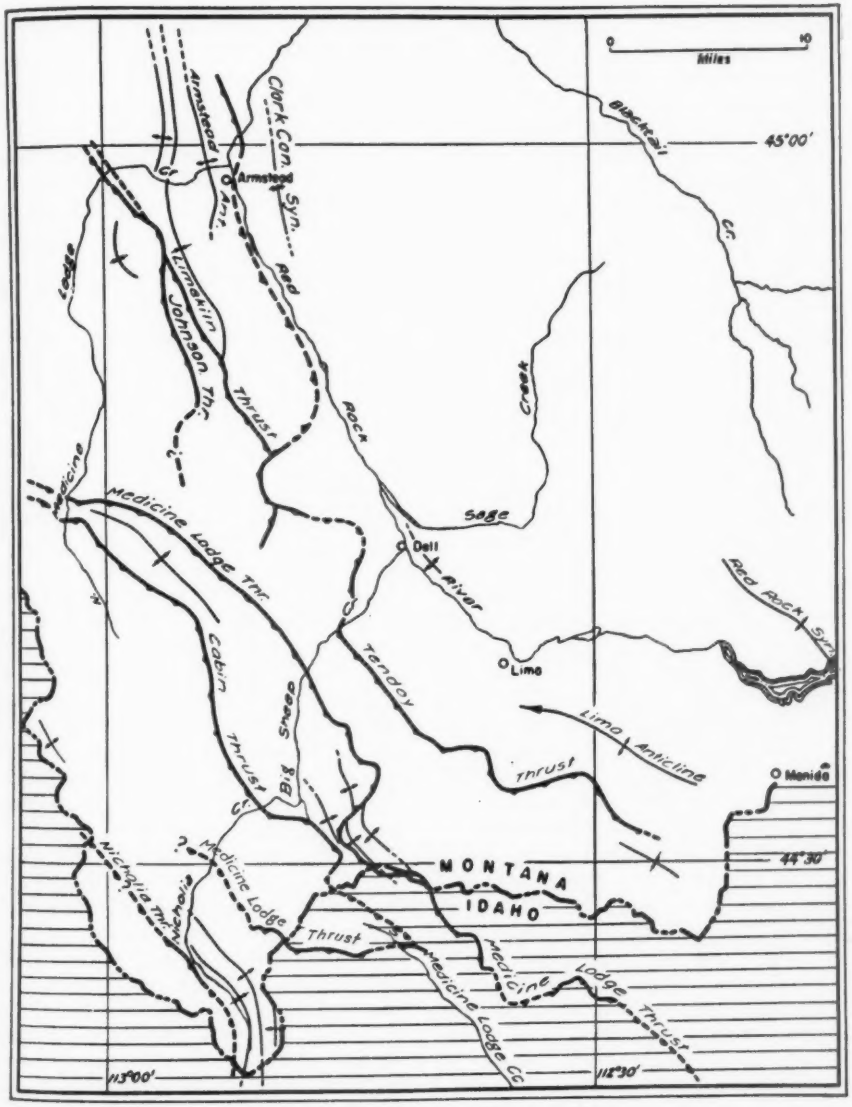

FIGURE 4.—LATE LARAMIDE TECTONIC MAP

response to thrusting. Several prominent and persistent folds with steep to overturned limbs occur in the Brazer strata east of Nicholia Creek. Beltian rocks are exposed in the center of a south-pitching anticline in the vicinity of Trail Creek. The Beaverhead granite intrudes this structure, and basalts overlap it. It is bounded on the southwest and southeast by high-angle faults and is believed to have been thrust in its entirety upon the Brazer (?) limestones to the northeast.

Other folds.—Ten miles south of Lima two anticlines with steeply dipping limbs and striking slightly north of east occur in the Beaverhead conglomerate. An identical trend was observed in the Beaverhead rocks south-

west of Kidd, where an intercalated freshwater limestone bed dips about 40°S. Erratically oriented dips between 20° and 35° are found in the Beaverhead east of the line Dell-Kidd. The relation of these various structures to the regional framework is not clear. They may be the result of erratic local forces during regional northeast-southwest compression.

Small, irregularly oriented folds are superposed on major anticlinal structures at the heads of Nicholia and Tex creeks. They are confined to the Brazer limestones. Most of them are chevron folds similar to small crumples in the Brazer strata of the Lemhi Range in central Idaho, described by Ross (1947, p. 1131). Ross interpreted them as due to erratic local forces applied under light load.

Low-angle thrusts.—Except for the northeast-striking thrust fault just beyond the eastern edge of the Lima region, thrusts in this part of southwestern Montana are confined to the area southwest of U.S. Highway 91 and trend northwest to north. The principal fault, the Medicine Lodge overthrust, was named by Kirkham (1927, p. 26), who observed that the Madison limestones along the State line near Irving Creek were thrust northeast across conglomerates which he believed to be Triassic but which actually belong to the Beaverhead formation of probable Paleocene age. The outcrop of the thrust can be traced from this point southeastward to the vicinity of Middle Creek Butte, except for a three mile interval where it is covered by Tertiary rhyolites. Southeast of Middle Creek Butte it disappears under the Snake River basalts of South Medicine Lodge Basin. Northwestward the thrust trace follows an irregularly winding course through the Tendoy Range in the area of Little and Big Sheep creeks. Madison and pre-Madison rocks overlie Permo-Triassic and Paleocene (?) strata (Pl. 4, fig. 2). Near Big Sheep Creek the thrust disappears under the Tertiary sediments of the down-faulted Muddy Creek Basin. The low-angle thrust emerging from beneath these beds farther northwest, in the area of McNinch Creek, is considered the continuation of the Medicine Lodge thrust. Here, the strike is slightly more westerly, and the stratigraphic throw has decreased, with Madison limestones overlying Quadrant quartzites. The thrust trace apparently continues under the Tertiary deposits along North Medicine Lodge Creek, beyond the western edge of the Lima region. In places the thrust plane is almost horizontal, as in the vicinity of the headwaters of Little Sheep Creek where the trace of the thrust makes several intricate loops. Elsewhere, dips up to 30° were found, such as near Irving Creek and near Two Spring Gulch.

Farther south, in the area drained by Divide and Deadman creeks, Brazer (?) limestones overlie Beltian and Kinnikinic quartzites. The contact dips gently south. The authors have tentatively interpreted this contact as a dislocated portion of the Medicine Lodge thrust. According to this hypothesis, the Medicine Lodge thrust was broken and the southern part tilted along the high-angle Cabin thrust. These relations are illustrated in cross section AA′A″A‴ and in the stereogram (Pl. 6). The northern and southern outcrops of the Medicine Lodge thrust converge to the southeast and the Cabin thrust probably merges with the southern portion somewhere under the Tertiary beds of South Medicine Lodge Basin. Beyond this point the Cabin thrust is believed to die out, leaving the unbroken Medicine Lodge thrust. To the northwest the southerly outcrop of the Medicine Lodge thrust disappears beneath the Tertiary deposits of Big Sheep Creek Basin.

The Medicine Lodge thrust is believed to have been similarly broken, displaced, and tilted by the high-angle Nicholia and Tendoy thrusts in the southern and central parts, respectively, of the Lima region (cross section AA′A″A‴, Pl. 6). The scattered outcrops of Madison limestone southwest of Lima, on the downthrown side of the Tendoy thrust, may represent small klippen of the Medicine Lodge thrust sheet overriding the Beaverhead conglomerate, or they may be autochthonous rocks in front of the Tendoy thrust, exposed by erosion of the Beaverhead. The rocks are too poorly exposed to permit a conclusion, but the autochthonous interpretation is shown in the cross section.

This structural interpretation of low-angle thrusting followed by transecting high-angle thrusting is partly hypothetical and requires

confirmation through additional detailed field work. It is believed, however, to provide the most satisfactory co-ordination of present data into a uniform structural framework.

The total horizontal displacement along the Medicine Lodge thrust is a matter of speculation. Kirkham (1927, p. 26) believes it to be several times 10,000 feet. One consideration is the pronounced difference between the geosynclinal facies of the rocks in the thrust sheet and the tectonic shelf facies in the autochthonous beds. Also, the fact that the Beaverhead conglomerate under the Medicine Lodge thrust sheet consists predominantly of Precambrian pebbles may be important. It suggests that, at the time of deposition of the conglomerate, the younger limestones had been stripped from a large area. The Mississippian limestones overlying these pebbles must have been brought in by thrusting from a considerable distance to the southwest. It would seem that the horizontal displacement must have been of the order of at least 10 miles. A similar figure is arrived at by speculation that the anticlinal crest in front of the Cabin thrust, near Four Eyes Canyon, was sheared off from its base north of Deadman Lake, 7 miles away (cross-section AA′A″A‴, Pl. 6). Allowing for some crustal shortening in the thrust sheet, the total displacement would be about 10 miles.

A second low-angle thrust, which may be a branch of the Medicine Lodge fault, is the Limekiln thrust. Displacement along it is also to the northeast. The thrust trace can be followed south-southeast through the Tendoy Range and is conspicuous in North Limekiln canyon. West of Armstead a klippe lies east of the main sheet. South of it the thrust plane splits into two parts, separated by a slice of Carboniferous rocks. Near Kelmbeck canyon the Limekiln thrust meets the high-angle Tendoy thrust. The relations between these two faults are not clear, but seemingly the Tendoy fault has broken through the Limekiln thrust sheet, cutting off a segment which was subsequently overridden by the Tendoy thrust sheet. Farther south, the Limekiln thrust is truncated by the high-angle normal fault along the eastern edge of Muddy Creek Basin, and on the downthrown side of this fault is covered by Tertiary sediments. Under these sediments it is probably linked with the Medicine Lodge thrust. In several places the Limekiln thrust is broken and displaced by high-angle faults. Madison limestones occur everywhere on the upthrown side of the thrust. Rocks on the downthrown side range from Mississippian to Cretaceous. Near South Limekiln Canyon the stratigraphic throw apparently is smallest, and Madison limestones occur on both sides of the thrust, which becomes difficult to trace. Local thrusting in the Madison can be observed in the Kelmbeck and South Limekiln canyons.

The Johnson thrust, named after Johnson Gulch, probably also belongs to this same phase of thrusting. Intense high-angle faulting has complicated the structural pattern of this thrust and the relationships are not completely understood. Northwest of Muddy Creek Basin pre-Beltian, Cambrian, Devonian, and Mississippian strata have been thrust over the Madison rocks of the Limekiln thrust. Farther northwest, near North Medicine Lodge Creek, a large area is interpreted as a window in the Johnson thrust. Much of this window is concealed by Cenozoic deposits. The Johnson thrust is cut by a large, northwest-striking normal fault. On the east, or upthrown, side of this fault the thrust sheet is represented by a large klippe of pre-Beltian rocks, resting on Madison and Amsden strata of the Limekiln thrust sheet (cross section JJ′, Pl. 6). The plane of thrusting appears to be undulating to some extent and possibly was folded. In general it is at a low angle to the horizontal. Southward, the trace of the Johnson thrust is lost in a maze of high-angle faults, but detailed work may show that the Johnson thrust is another branch of the Medicine Lodge overthrust.

High-angle thrusts.—The high-angle Nicholia thrust was postulated to explain the juxtaposition of Beltian and Mississippian (Brazer ?) rocks on opposite sides of Nicholia Creek Valley. No actual outcrop was found, for glacial drift conceals the bedrock on the valley bottom. The thrust strikes approximately northwest and is believed to dip southwest. A high angle of dip is indicated by the absence of Belt rocks on the east side of the deep and narrow valley, where the Brazer (?) limestones

make a steep wall. To the northwest the outcrop of the thrust is concealed by Tertiary deposits in Big Sheep Creek Basin. To the southeast it apparently is truncated by a transverse high-angle fault, possibly a tear fault. Farther south, along the Idaho border, a klippe of Beltian and Ordovician rocks is superposed on Madison limestones. It is believed that the fault under this klippe is a part of the Nicholia thrust. The fault truncating it on the southwest probably is another tear fault.

Another high-angle fault was named after Cabin Creek, a tributary entering Big Sheep Creek from the west. Along its known length the Cabin thrust brings rocks of the Precambrian Dillon granite gneiss on top of Madison limestones, which locally are highly contorted. It can be traced for many miles parallel to the southwest margin of the Tendoy Range but is truncated in several places by the Deadman fault which bounds the range. The downthrown portion of the thrust plane is concealed beneath Tertiary beds in Big Sheep Creek Basin. Southeastward, across the State line, the thrust trace on the pre-basin bed surface is believed to continue to a point where it joins the southern outcrop of the Medicine Lodge overthrust, and to die out a short distance beyond (stereogram, Pl. 6). On the upthrown side of the Deadman fault soil and vegetation cover the thrust trace, but its course appears to be relatively straight revealing the steep dip of the fault plane. An exposure in a gulch near the northwestern extremity of the thrust reveals a dip of 60° SW.

A third high-angle thrust is evident near the northeastern edge of the Tendoy Range and has been termed Tendoy thrust. The northernmost exposure of this thrust is in the area north of Armstead, where Madison limestones overlie the Paleocene (?) Beaverhead formation. In Red Rock Basin Cenozoic deposits cover the thrust trace as far south as Dell, except in the area of McKnight Canyon, where the thrust makes a sharp loop to the west into the Tendoy Mountains, apparently cutting off a portion of the earlier Limekiln thrust. Reappearing on the upthrown side of the Red Rock fault zone west of Dell, the Tendoy thrust continues southeastward through the mountains and parallel to the mountain front, with locally highly contorted Madison along most of the upthrown, and Beaverhead on the downthrown side. Southeast of Lima the stratigraphic throw gradually decreases as successively younger rocks appear in the thrust sheet along the fault trace. Aspen strata in the central part of the Lima anticline appear in the autochthonous rocks. Finally, Aspen beds occur on both sides of the thrust, and it becomes impossible to trace it farther southeast. The thrust appears to die out in this vicinity. The thrust plane dips steeply; it is well exposed 5 miles south of Lima, where Quadrant quartzites rest on Beaverhead conglomerates. At this place the dip is about 60° SW.

Other Laramide faults.—The principal persistent normal faults in the Lima region are post-Laramide and strike consistently northwest. However, a number of smaller high-angle faults with diverse trends seem to be genetically related to the Laramide folds and thrusts. In the southwestern corner of the region two northeast-trending faults west of Nicholia Creek have been mentioned as probable tear faults associated with the Nicholia thrust. A third fault in this vicinity, which parallels the State line, brings Kinnikinic (?) quartzites in contact with Madison limestones. It appears to be cut by the Beaverhead intrusive and to disappear northward under volcanic rocks of probable mid-Tertiary age. This fault probably is associated with the Laramide orogeny. Farther north, near Simpson Creek, an east-west fault separates Quadrant from Brazer rocks.

In the north-central Tendoy Range, north of Muddy Creek Basin, the short, irregular high-angle faults associated with the Johnson thrust probably were formed in response to the thrusting. The same is true for the east-northeast-trending tear faults that break through the Limekiln thrust south of Armstead. Small cross faults in the east flank of the Armstead anticline also may be Laramide.

A number of high-angle faults are concentrated in areas where northwest structures were superposed on pre-existing northeast trends, and apparently are the result of the stresses created during this superposition. They are found in the Little Water syncline south of Kidd, in the southeast flank of the

Garfield anticline south of Lima, and in the Clover Creek anticline in the Snowcrest Range. One small fault is present where the northwesterly extension of the Red Rock syncline intersects the southeast limb of the Blacktail-Snowcrest anticline. These faults have scattered orientations. They are responsible for offsetting conspicuous ridges of Thaynes, Rierdon, and Kootenai strata as much as 750 feet. Both normal and reverse faults are present in the Garfield anticline. Rotation took place in several blocks, and the composite crustal movements apparently were complex.

Several normal faults in the vicinity of the Lima Reservoir involve the Beaverhead conglomerate and Aspen formation. They trend easterly to northeasterly, and are clearly traceable on aerial photographs; they are tentatively considered to be Laramide, for their trend is normal to the southeast-northwest pattern of normal faults known to be post-Laramide.

Post-Laramide Structures

General remarks.—Post-Laramide structural elements in the Lima region include gentle folds and flexures and normal high-angle faults. The latter include large block faults, which bound the topographic basins and produced tilting of the basin sediments. They trend northwest to north. Smaller high-angle faults with similar trends are found in the region.

For most of these structures a post-Laramide age is indicated by the fact that they involve mid-Tertiary and late Tertiary basin beds and truncate Laramide structures. Some of the faults have been active in Recent time, but whether movements had already been initiated prior to deposition of the basin beds and were responsible for creating the basins was undetermined.

Folds.—A north-south-trending anticline exists in the beds of the lower Miocene Blacktail Deer Creek formation in the vicinity of the West Fork of Blacktail Creek. Dips on either flank range from 5° to 10°. No younger sediments occur in this area; deformation must be post-lower Miocene.

Persistent flexures occur in several places in the Medicine Lodge beds, and north-south folds are in evidence near Kissick Canyon. In the area of South Medicine Lodge Creek the beds are almost horizontal. Hayden (1872, p. 147–149) reports a gentle anticline whose axis approximately follows the creek, but this observation may have been made farther southeast, outside the limits of the Lima region. Locally, sharp reversals of the predominantly northeasterly dip of the Medicine Lodge beds are found near the Tendoy mountain front. These may be ascribed not to folding, but to tilting of minor fault blocks.

Block faults.—There are indications of an important block fault, here termed Deadman fault, along the southwest front of the Tendoy Range. Along this front Precambrian rocks rise high above the adjacent basin, and the escarpment shows remnants of triangular facets. Where Deadman Creek cuts through the mountain front a fault breccia is exposed, with angular fragments of Dillon granite gneiss. Steeply dipping slickensided and polished shear planes in this vicinity strike approximately parallel to the mountain front. Along the mountain front near the Idaho State line a conspicuous cliff of Medicine Lodge freshwater limestone is veneered by Edie School rhyolite. As the main body of these lavas occurs in the adjacent South Medicine Lodge Basin, this cliff must be a remnant of a former fault scarp, displaced northward by weathering and erosion.

Additional proof of a continuous fault or fault zone along this mountain front is afforded by the almost uniform northeasterly dip of the Medicine Lodge beds, evidently produced by tilting. Dips as high as 35° NE. occur adjacent to the mountains and could hardly be depositional. No fault is in evidence from Black Canyon to the southeast, and in this areas the basin beds are nearly horizontal. To the northwest the Deadman fault dies out north of Island Butte. The total length of the fault exceeds 20 miles.

The exact period of activity along the Deadman fault is in doubt. Late Tertiary or Quaternary movements displaced the Edie School rhyolites more than 500 feet, but probably movement started earlier and was partly

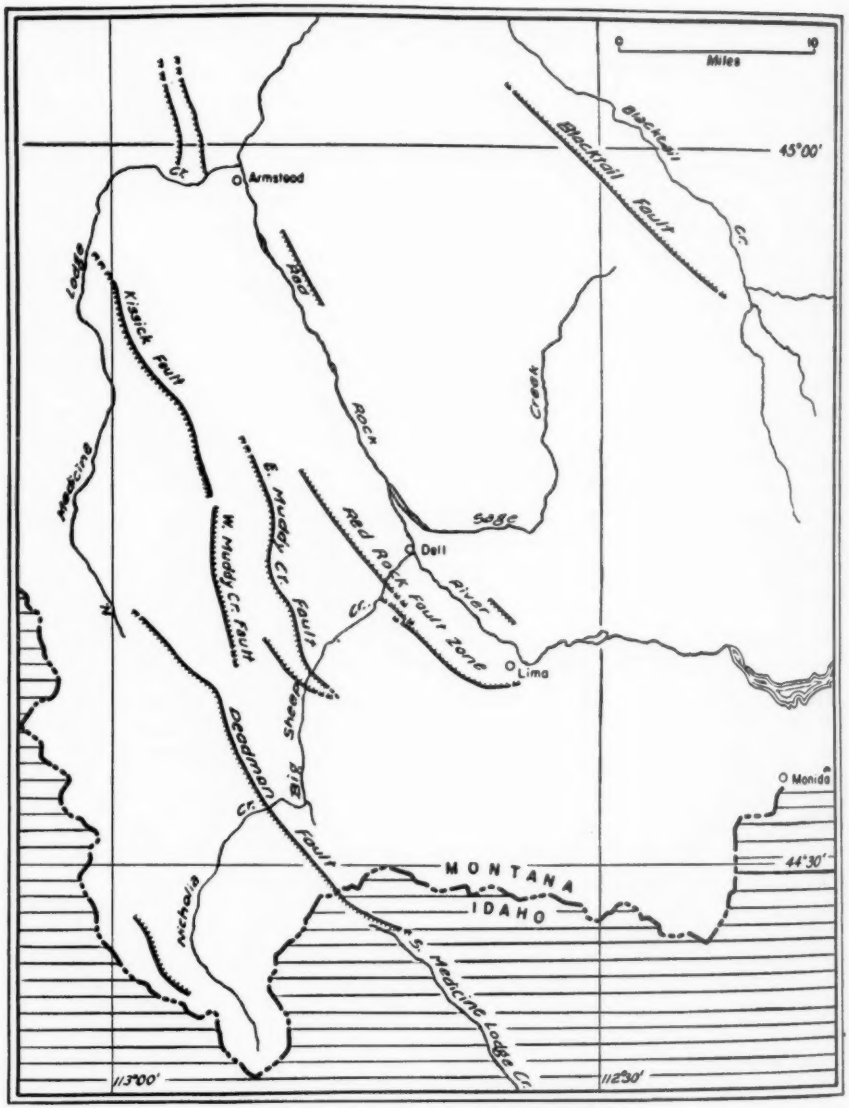

FIGURE 5.—POST-LARAMIDE TECTONIC MAP

responsible for creating the depression in which the Miocene Medicine Lodge beds were laid down. The aggregate displacement may amount to several thousand feet.

The Kissick fault is another high-angle normal fault bounding the Tendoy Range on the west. Along part of its course it separates Tertiary beds of North Medicine Lodge Basin from older rocks in the Tendoy Range. Elsewhere, it cuts through the Precambrian rocks which override Madison limestones along the Johnson thrust. The displacement along this fault is unknown. The fault was active at least as late as Miocene, as indicated by the prob-

able Miocene age of the Medicine Lodge beds. Initial movements may have been earlier.

Muddy Creek Basin, an intramontane depression in the Tendoy Range, is bounded on east and west by normal faults, making the basin a structural graben. The West Muddy Creek fault forms a distinctive scarp, separating Madison limestones from the Tertiary (Oligocene ?) Muddy Creek basin beds and volcanics, except east of Graphite Mountain, where the basin beds extend farther west and no evidence of faulting could be found. Either the fault dies out in this vicinity, or the displacement becomes too small to bring up pre-Tertiary strata on the west side. East Muddy Creek fault is a distinct feature, extending persistently along the entire east side of the basin for more than 15 miles. It truncates the Laramide Little Water syncline. A smaller normal fault bounds Muddy Creek Basin on its southern extremity, dying out in basin beds in both directions. It crosses Big Sheep Creek canyon, where alluvium conceals it. The maximum displacement along the Muddy Creek faults is unknown, but the general eastward dip, locally as high as 35°, of the Tertiary basin beds indicates that the vertical throw was greatest along the eastern structure.

A series of en echelon normal faults along the eastern front of the Tendoy Range constitutes the Red Rock fault zone. A steep, composite fault scarp with excellently preserved triangular facets characterizes the front of the range between Lima and Kidd. The escarpment truncates Laramide structures. Recent movements along the faults are indicated by scarplets cutting through modern alluvial fans spreading eastward into Red Rock Basin from the heads of canyons (Pl. 4, fig. 3). Locally, small fans have been built on top of the large ones. Historic earthquakes probably are attributable to further adjustments along this fault zone.

The fault zone curves eastward near Lima and apparently dies out there. To the northwest the fault zone can be traced as far as McKnight Canyon. Possibly easterly offsets continue under the Quaternary valley fill to, or beyond, Armstead. Judging from the height of the triangular facets above the basin surface, the vertical throw is at least 500 feet, but the cumulative displacement probably is much greater. Pardee (1950, p. 377) considers that the entire mountain front probably is a badly worn fault scarp and that the total vertical displacement may be as much as 3000 feet.

The steep northeast flank of the Blacktail Range is an escarpment associated with another major block fault. Nearly 15 miles long within the Lima region, the Blacktail fault extends at least several miles beyond its northern limit. The fault is postulated on physiographic evidence only. Its present trace is covered by Quaternary fanglomerates. Pardee (1950, p. 385) also recognizes the fault. It cuts through and displaces the exposed Precambrian core of the Blacktail-Snowcrest anticline, striking normal to its axis. Farther southeast the Blacktail fault cuts through the Miocene Blacktail Deer Creek formation and dies out in these beds. In the vicinity of the fault the Tertiary beds are tilted to the southwest.

The amount of vertical throw is unknown, but from the height of the escarpment it appears to have been at least 2500 feet near Ashbough Canyon. Streams flowing from the mountain front have incised their courses 50 to 100 feet into the fanglomerate deposits, indicating no recent movements. Small fans now building at the mouths of some of the canyons may be the result of overgrazing along the headwaters of the streams. Pardee (1950, p. 386) believes that much of the movement along the fault occurred within the Pleistocene.

Other post-Laramide faults.—A number of smaller faults were observed or postulated in the Lima region. Not all of these have been definitely established as being post-Laramide, but they are tentatively considered so because of their normal nature and northwest trends, which fit them into the tectonic framework of the known post-Laramide faults.

Three such faults bound the Beaverhead Range on the northeast, near Simpson Creek. In the Tendoy Range a small northwest-striking fault was postulated near the Idaho border, east of Bannock Pass. Warm Creek derives its name from a group of warm springs which feed the creek along this fault. The fault separates Medicine Lodge from Madison limestones and is truncated by an erosion surface, probably late Tertiary. Two normal faults parallel the Kissick fault to the east, and two

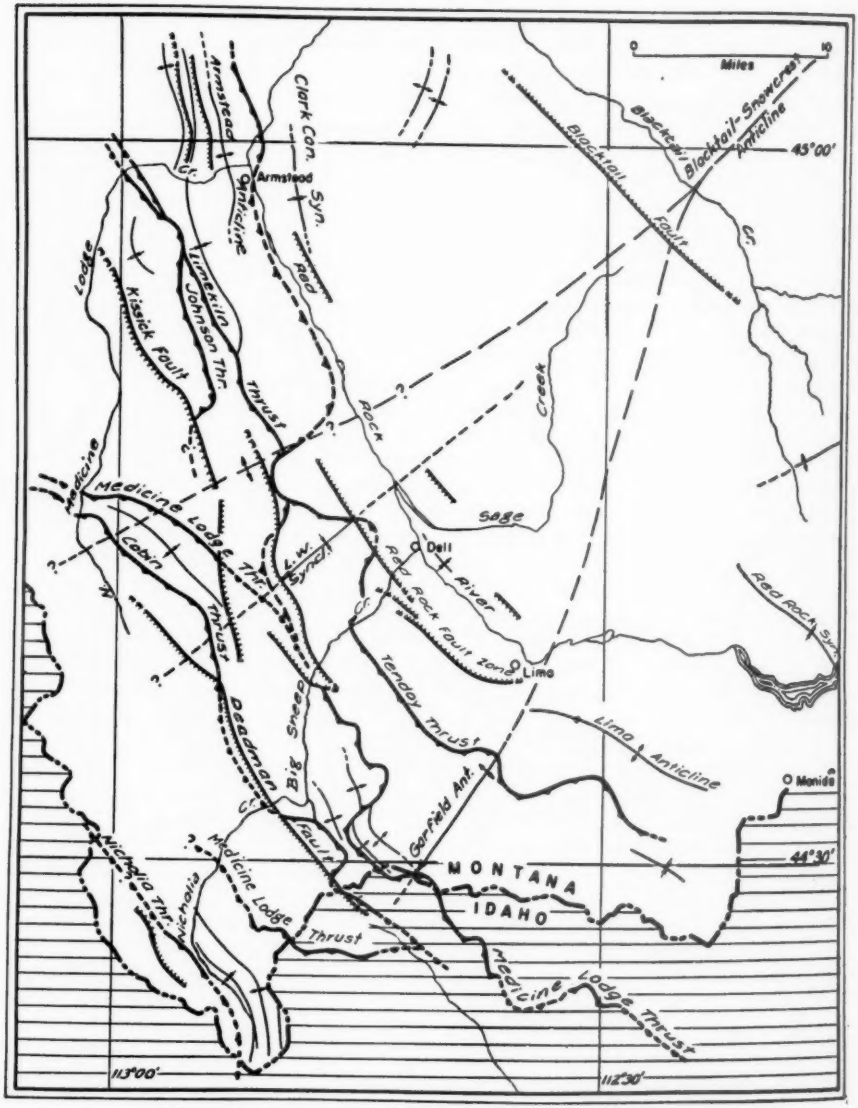

FIGURE 6.—COMPOSITE MAP OF CENOZOIC TECTONIC ELEMENTS

others with more northerly trends appear in the west flank of the Armstead anticline. The youngest beds cut by these structures are Cretaceous.

Discontinuous normal faults occur along the east side of Red Rock Basin, between Lima and Armstead. All except one are downthrown on their southwest sides, making that basin an imperfect graben. Some of these faults are quite long; the one nearest Armstead extends about 7 miles, but displacements are minor. The rocks affected include the Beaverhead conglomerates and the Sage Creek, Cook Ranch, and Blacktail Deer Creek formations. Hence, Miocene or

post-Miocene movements must have taken place. Several lesser faults parallel the Blacktail block fault in the vicinity of Blacktail Creek and are related to this major structure.

Regional Relations

The thrusts in the Lima region fit into a regional pattern of northwest- to north-trending overthrusts with eastward displacements in the northern Rocky Mountains. West of the Lima region the Lemhi-Hawley, Lost River, and other thrusts of central Idaho were mapped by Kirkham (1927, p. 27–29) and Ross (1947, p. 1133–1135). According to Kirkham several thrust sheets moved west, but this conclusion was not substantiated by Ross. The eastern limit of this major thrust belt is the Tendoy thrust. The belt continues far beyond the northern limit of the Lima region, but how far the individual structures persist is not known. Southeastward, several of the thrusts die out or unite, but the thrust belt appears to continue beneath the lavas of the Snake River plains of central Idaho, where regional downwarp occurred in late Cenozoic time (Kirkham, 1927, p. 13, 24–26; 1931, p. 467–482). The thrust belt re-emerges at the southeastern end of this downwarp and continues in the great overthrusts of western Wyoming and southeastern Idaho.

Kirkham (1927, p. 27) has speculated that the Medicine Lodge thrust may be a northerly extension of the Bannock overthrust of southeastern Idaho. If the outcrop of the Bannock thrust as mapped by Mansfield (1927, p. 150–151) is projected about 80 miles along strike from the point where it disappears under the Snake River basalts, it reaches the north side of the plains about where the Medicine Lodge thrust emerges from beneath the same lavas. Carboniferous beds occur on the upthrust sides of both thrusts. With an exposed length of 270 miles for the Bannock thrust, a known length in excess of 50 miles (and possibly much greater) for the Medicine Lodge thrust, and an intervening distance of 80 miles, the structure, if actually continuous, would be one of the largest and most persistent in the Cordilleran region.

The post-Laramide block faults likewise form part of a regional belt, which coincides approximately with the thrust belt, although at least one important fault, the Blacktail, occurs farther east. West of the Lima region block faults have been reported to bound the southwestern margins of the Beaverhead and Lemhi ranges (Anderson, 1934, p. 19, 21; 1947, p. 67–68; Ross, 1947, p. 1137–1139). Lengthwise, this belt of trenches can be traced for 2000 miles through the Rocky Mountain region and has been compared to a rift zone (Eardley, 1947, p. 1176; 1951, p. 494). The block faults have a profound influence on the modern topography.

Geomorphology

Mountain Ranges

The straight steep slopes and parallel northwest trends of the Blacktail and Tendoy ranges and the adjacent intermontane and intramontane basins are the results of the major post-Laramide block faults. Large alluvial fans flanking the northeast mountain fronts of these ranges are also related to the faults (Pl. 4, fig. 3). No large block faults bound the Beaverhead Range on the northeast side, which consequently is much less regular and abrupt. The northeast trend of the Snowcrest Range is clearly due to the strike of the Paleozoic and Mesozoic strata, whereas the trend of the Ruby Range appears to be largely controlled by structural patterns in the Precambrian rocks which constitute its core.

Stream Valleys

The major trunk streams are largely controlled by the regional basin and range pattern. In the intermontane basins these streams flow northwest or southeast (Fig. 2). Examples are the Red Rock River and Blacktail, Muddy, North Medicine Lodge, and South Medicine Lodge creeks.

Several creeks leave the basins to re-enter the mountains in narrow canyons. A most striking example is the Nicholia-Big Sheep Creek system. The headwaters of Nicholia Creek and its tributaries are on the northeast slope of the Beaverhead Range. The stream then crosses Big Sheep Creek Basin and enters

a gorge in the Tendoy Range, at which point the stream is called Big Sheep Creek (Pl. 3, fig. 2). The gorge cuts through the range in northeasterly direction, normal to the strike of the beds. Crossing the intramontane Muddy Creek Basin the stream does not follow this basin longitudinally to leave it farther to the northwest through a low and easily erodable divide at the head of Little Water Creek. Instead, it continues northeastward through a deep gorge cut into highly resistant quartzites of the Tendoy Range.

Deadman and Cabin creeks also cut through parts of the Tendoy Range along their lower courses. Horse Prairie Creek has cut across the mountains and its valley separates the Tendoy Range from the mountains north of Armstead.

These abnormal characteristics of the modern drainage system cannot be explained by normal headward erosion, as was done by Umpley (1913a, p. 29–30). Headward erosion through the soft Mesozoic strata at the head of Little Water Creek should take far less time than through the much harder and more extensive Paleozoic quartzites along Big Sheep Creek. Consequently, if only headward erosion had taken place, Muddy Creek Basin would now be drained northeastward via Little Water Creek into the Red Rock River, rather than southeastward by Muddy Creek into Big Sheep Creek.

Rather, Big Sheep Creek canyon must be regarded as antecedent to the faults along which the present Tendoy Range was uplifted. West of the Lima region the abandoned bed of a late Cenozoic stream ("Arco River") has been traced in northeasterly direction through wind gaps high in the Lost River and Lemhi ranges (Anderson, 1947, p. 70). The probable northeast continuation of this river can be traced across part of the Lima region. In line with the two wind gaps occurs another wind gap at the head of Trail Creek in the Beaverhead Range. This gap is a channel in a high-level erosion surface covered by old stream gravels. Farther northeast remains of the Arco River system can be found in the Tendoy Range where old river gravels are preserved on a similar high surface. In places these gravels are water-rounded pebbles and boulders of Precambrian (Beltian) quartzite. As these quartzites are exposed in place only in the Beaverhead Range, their occurrence as pebbles in the Tendoy Range is further support for a northeast drainage on the high surface.

The Big Sheep Creek system, including Trail Creek, Nicholia Creek, Deadman Creek, and Cabin Creek, is believed to have been inherited from the lower course of this ancient Arco River system. Downward erosion along these streams kept pace with the late Cenozoic rise of the Tendoy Range along major block faults. The Beaverhead, Lemhi, and Lost River ranges, however, rose too rapidly for erosion to keep pace, and the Arco River was separated and became dry along the present wind gaps. Barbed tributaries on the west side of the Beaverhead Range indicate a reversal of flow (Anderson, 1947, p. 71–73).

Another anomalous stream is South Medicine Lodge Creek, which flows through the intermontane basin between the Tendoy and Beaverhead ranges. In contrast to other basin streams it occupies a deep canyon in upper Tertiary limestones and overlying lavas. Cutting of this canyon probably was related to the regional upwarping of an east-west trending arch in southwestern Montana and central Idaho, complementary to the Snake River downwarp farther south (Kirkham, 1927, p. 13, 24–26; 1931, p. 467–82). The axis of this arch coincides with the topographic axis of the Centennial Range, and Kirkham thinks that its westerly continuation is the cause of the low divides in the basins adjoining the Beaverhead, Lemhi, and Lost River ranges. The Bannock and Monida passes, both on the Continental Divide on either side of the Tendoy Range, are in line with this trend. Stream rejuvenation was most pronounced on the south side of this axis where the erosion base in the Snake River region had been greatly depressed. This explains the canyons of South Medicine Lodge Creek and its tributaries and the absence of similar canyons in the basins north of the arch axis.

Some of the tributaries to the main streams are adjusted along parts of their courses to the Laramide structures and to the lithologies of the underlying rocks (Pl. 1). One example is the Middle Fork of Little Sheep Creek, whose upper course is subsequent, following soft Triassic and Permian beds and making the same sharp

bend to the north. The East Fork of Little Sheep Creek and the upper course of Deep Creek are subsequent also, following the Laramide trends of the beds. These streams are fed by numerous obsequent and resequent gulches which contribute to a distinct trellis pattern in places.

Erosion Surfaces

General remarks.—The present topography of the Lima region is the result of three main erosion cycles, interrupted in various stages of advancement. The result is a series of three main erosion surfaces in different stages of development. In the following discussion the post-Laramide, mid-Tertiary surface is called "old surface", the late Tertiary surface is called "intermediate surface", and the compound Quaternary surface is called "young surface".

Old surface.—The relatively uniform altitude of ridge crests in the Beaverhead Range and the gently rolling crests of the Tendoy and Blacktail ranges suggest the existence of summit erosion surfaces 9000 to 9500 feet above sea level (Pl. 5, fig. 1). In the Tendoy Range the summit erosion surface is not younger than mid-Tertiary, as a surface shown to be of late Tertiary age occurs at a considerably lower level, and a chain of events separates the development of these two surfaces. Several high peaks rise above the Tendoy summit level.

The Beaverhead summit level may correlate with the mid-Tertiary surface in the Tendoy Range. However, it is not clearly differentiated from a high-level surface covered by late Tertiary or Quaternary gravels, and may have developed only in late Tertiary time. The same statement applies to the Blacktail Range summit level. On the northeastern slopes of the Beaverhead Range the mid-Tertiary Medicine Lodge lavas occupy ancient valleys. The floors of these valleys must represent the old surface. This indicates that the original surface was not a peneplain, but had a moderate, perhaps late mature, relief. A similar statement was made by Pardee (1950, p. 366) concerning a mid-Tertiary erosion surface postulated for western Montana.

In the intervening basins, the floors beneath the mid-Tertiary sediments probably represent the same old surface depressed by crustal disturbances. This diastrophism may have been crustal warping, but very possibly down-faulting along the major block faults bounding the ranges was initiated during this time. Pardee (1950, p. 366) considers crustal disturbance responsible for disrupting the old surface in western Montana.

Alternatively, the basins may have been due to rejuvenated erosion which carved depressions in the old surface, but it would be difficult to understand, in many places, why these basins were carved in metamorphic Precambrian rocks, leaving the softer limestones to stand out as mountain ranges. Such is the relationship between the Beaverhead Range and parts of North Medicine Lodge and Big Sheep Creek basins, as well as between Big Sheep Creek Basin and parts of the Tendoy Range (Pl. 1).

Creation of the mid-Tertiary basins by block faulting would also account for their linear, northwest trend, as indicated by the present distribution of the basin sediments. However, in the eastern half of the Lima region, where block faults are less prominent and where the basins beds are not confined to northwest belts, the depressions may have been produced by warping or by erosion rejuvenated through regional uplift. The new erosion cycle was responsible for the intermediate surface.

Intermediate surface.—Remnants of an erosion surface conspicuously below the summit level (at 8000–8500 feet) are well developed in the Tendoy Range (Pl. 5, fig. 2), where they are preserved as small plateaus, bench levels, and flat-topped ridges. These surfaces are covered by numerous patches of gravel. Gravel deposits also cover high surfaces in the Beaverhead Range. No high-level gravels are found in the Blacktail Range.

The youngest rocks underlying this surface are the limestones of the Miocene Medicine Lodge basin beds. These are preserved on a small plateau high in the Tendoy Range, northeast of Bannock Pass (Pl. 5, fig. 2). Similarly, limestones of the Miocene Blacktail Deer Creek beds underlie a flat plateau at an altitude of about 8500 feet north of the Snowcrest Range, in the vicinity of the West Fork of Blacktail Creek (cross section K, K′, K″, K‴,

Pl. 6). This indicates a late Tertiary to early Quaternary age for the intermediate erosion surface.

A lower surface of late Cenozoic age occurs in the adjacent basins and valleys, where it cuts across mid- to upper Tertiary sediments and lavas. It can be observed in the Sage Creek area, in Muddy Creek Basin, in North and South Medicine Lodge basins, and in Big Sheep Creek Basin. A low divide separating the latter two basins at Bannock Pass probably is younger.

In Red Rock Basin this same surface probably is present below the thick Quaternary cover, but to the east it rises, partly along faults, and cuts across lower Tertiary Beaverhead conglomerates between Armstead and Dell. Where exposed, this surface is 6500 to 7500 feet above sea level.

Evidence in the Bannock Pass area proves that this lower late Cenozoic surface once was continuous with the late Cenozoic surface in the Tendoy Range. A patch of late Tertiary (Edie School) rhyolite lava is preserved on a small plateau in this range. Some 500 to 1000 feet lower, the same lava overlies the flat surface on the basin beds along South Medicine Lodge Creek (cross sections EE′, GG′, Pl. 6). In both places thin gravel patches overlie the lava. The Deadman block fault separates the two surfaces. Evidently, an erosion surface once extended uninterruptedly across the region of the Tendoy Range and the basin to the southwest, but was broken and dislocated along block faults in latest Tertiary or early Quaternary time.

The writers believe the Muddy Creek basin surface and the floor below the Quaternary cover of Red Rock Basin are additional portions of the same erosion surface, downfaulted along the Muddy Creek faults and the Red Rock fault zone. The floor of Blacktail Creek valley may occupy a similar relation to the Blacktail fault, which bounds the adjacent range; it is also covered by Quaternary deposits. The high-level representatives of the intermediate surface in the Blacktail and Beaverhead ranges and north of the Snowcrest Range are considered to have reached their present levels as a result of the warping, or faulting and tilting which lifted these ranges.

Large parts of the intermediate surface were and are essentially flat, but these parts were flanked by hilly areas upholding remnants of the older erosion surface. Elsewhere, notably in the Beaverhead and Blacktail ranges, the two surfaces may have been essentially coextensive, or the old surface may have been destroyed.

The extensively gravelled portion of the intermediate surface in the Tendoy and Beaverhead ranges and the intervening area are believed to represent the bedding of the ancient Arco River.

Young surface.—A third surface is represented by terraces and modern stream flats along some of the major drainage lines. It is therefore a compound surface and is depositional as well as erosional. Its altitude ranges from about 7000 feet in Big Sheep Creek Basin to about 5500 feet in Red Rock Basin near Armstead.

Two river terraces can be observed along Nicholia Creek in Big Sheep Creek Basin; in North Medicine Lodge Basin and Muddy Creek Basin; in the Sawmill-Junction Creek area southeast of Lima; along the Red Rock River; and elsewhere. The terraces probably are of Pleistocene age and related to near-by mountain glaciation. No terraces are present in South Medicine Lodge Basin where the major streams occupy canyons.

The mountain ranges have acquired an early mature topography. Some of the larger canyons have developed flat bottoms and are filled with alluvium, as along Big Sheep Creek (Pl. 3, fig. 2).

Glacial Features

Erosional features.—Glacial cirques occur at the heads of several major valleys in the Beaverhead Range. Especially well developed are those of Nicholia and Meadow Creek canyons. The continental Divide at the head of Nicholia canyon is extremely sharp. A few Matterhorn peaks developed along the Divide. Good examples are Italian Peak at the southernmost tip of Montana, and Eighteen Mile Peak, farther northwest.

Nicholia canyon is U-shaped, and its steep walls are largely covered by talus. Tributary

valleys are at grade, and the main valley evidently was not deepened much by the glacier.

Less conspicuous cirques and U-shaped valleys are found in the Tendoy Range at the head of Irving Creek.

Depositional features.—Two superimposed morainic deposits are found in and adjacent to the Beaverhead Range. The older drift extends farther eastward and northward into Big Sheep Creek Basin. It is characterized by low rolling hills which blend into the highest stream terrace. Glacial lakes have been filled, and glacial boulders are about two-thirds buried in the clay matrix.

Younger drift covers the older moraine in Nicholia canyon and other valleys and in the basin immediately adjacent to the mountains. It has been affected only slightly by erosion and has preserved a hummocky topography. Several glacial lakes persist, and boulders are scattered across the surface. Excellent terminal and recessional moraines are preserved at the mouth of Nicholia canyon, below the junction of Trail Creek canyon (Pl. 5, fig. 3).

One moraine was observed along McNinch Creek in the Tendoy Range west of Muddy Creek Basin. Apparently, glaciation was mostly confined to the Beaverhead Range, which was probably higher than the remainder of the Lima region.

Regional Relations

The linear basin and range topography in the western part of the area fits into a regional pattern of similar parallel features. To the west, in central Idaho, the Salmon River, Lemhi, and Lost River ranges trend parallel to the Beaverhead and Tendoy ranges and, like the latter, plunge southeastward under the lavas of the Snake River plains. These ranges are also separated by wide, flat-bottomed basins. Drainage through these basins is to the southeast and northwest, similar to South and North Medicine Lodge creeks.

This distinctly linear basin and range pattern has prompted Anderson and Wagner (1944, p. 4) and Ross (1947, p. 1140) to ally the region with the Middle Rocky Mountain physiographic province, though it is separated from it by the wide expanse of the Snake River plains. Fenneman (1930) included the region in the Northern Rocky Mountain province.

A high erosion surface, represented by the summit levels of the Salmon River, Lost River, Lemhi, and Beaverhead ranges, has been variously designated as "summit peneplain" (Atwood, 1916, p. 707), "Idaho peneplain" (Mansfield, 1924, p. 472; 1927, p. 354) and "summit erosion surface" (Anderson, 1947, p. 63, 65). The concept that the basins are depressions in the old summit level and that therefore basin floors and summit level were once coextensive, has been defended by Umpleby (1912, p. 42; 1913b, p. 226–227), Atwood (1917, p. 545), Kirkham (1927, p. 11–12) and Perry (1934, p. 6). Others have correlated the summit level with the surface on top of the Tertiary beds in the basins (Blackwelder, 1917, p. 545; Rich, 1918, p. 135–136). Ross (1937, p. 88–89; 1947, p. 1141–1144) believes that the old erosion surface correlates with a surface of post-mature relief cut across the mid-Tertiary Challis volcanic rocks in parts of central Idaho.

Evidence from the Lima region indicates that the surface across the basin beds is the equivalent of the intermediate (late Tertiary) erosion surface in the Tendoy Range. As the summit level of that range is higher and therefore earlier, it is probably coextensive with the basin floor. Kirkham (1927, p. 13) described an intermediate erosion surface from a large region in central Idaho. It is probably correlative with the Donkey Hills surface in the Borah Peak quadrangle of central Idaho. (Ross, 1947, p. 1145–1146). Pardee (1950, p. 366) has observed a late Tertiary erosion surface in many parts of western Montana.

A composite low surface comparable to that observed in the Lima region has been described from the Centennial region farther east (Honkala, 1949, unpub. thesis, Univ. Mich.).

Finally, glacial features in the Lima region are provisionally correlated with the Bull Lake and the younger Pinedale glaciations of the Wisconsin stage as recognized in Wyoming (Blackwelder, 1915, p. 310).

Geologic History

Precambrian Events

The earliest traceable events in the history of the Lima region were the deposition of the sediments and the intrusion of the igneous materials of the Pony group. These were subjected to orogenic movements, metamorphosed to gneisses, and partially eroded before deposition of the sediments of the Cherry Creek group. Another major orogeny transformed the Cherry Creek rocks into gneisses, schists, and silicate marbles. In the late stages of this orogeny the Dillon granite batholith intruded and altered the pre-existing metamorphic rocks, causing serpentinization of the silicate marbles. The final phase of the orogeny produced gneissic textures in the main mass of the batholith, but local post-tectonic intrusions escaped foliation.

After a long period of erosion had reduced the land to a low elevation, the graywacke sediments of the Belt group were laid down in shallow seas or on adjacent flood plains in the southwestern part of the Lima region. Tectonically, this area was part of the Rocky Mountain miogeosyncline during this period. It is not known whether the Belt sediments extended also across the northern and eastern parts of the Lima region, where no deposits of that age are found today. A westerly derivation of the Montana Belt materials is indicated by a regional coarsening and a decrease in nonclastic constituents in this direction (Schuchert and Dunbar, 1941, p. 65–66). The abundance of feldspar (K feldspar and andesine) in the Belt rocks of the Lima region suggests that at least part of the material was derived locally from the Pony gneisses and the Dillon granite gneiss. Precambrian time ended with regional uplift without significant folding.

Paleozoic Events

Erosion during the early Cambrian removed any Belt strata that may have been present in the eastern and northern parts of the region. In Middle Cambrian time a shallow sea transgressed the area. Sediments derived chiefly or wholly from the east were deposited here, beginning with the conglomerates and quartz sands of the Flathead formation, eroded from the landmass of Precambrian rocks. This was followed without interruption by deposition of the Middle Cambrian Wolsey shale, Meagher dolomite, Park shale, and the Upper Cambrian Pilgrim dolomite on the tectonic shelf east of the geosyncline.

Uplift and erosion during the Late Cambrian and the Early and Middle Ordovician removed all Cambrian deposits from the southwestern part. This area was flooded again in the late Ordovician, and the clean quartz sands of the Kinnikinic formation were laid down. The sand probably represents a second- or third-cycle deposit derived from the Precambrian and Cambrian rocks and, according to Ross (1934, p. 997), deposited on shoal flats at the mouths of large rivers. Kinnikinic sedimentation in the Lima region took place considerably east of the axis of the Paleozoic geosyncline, which had become increasingly well defined.

If Silurian and Lower and Middle Devonian sediments were laid down, they were removed prior to the Late Devonian, when the Jefferson dolomites were deposited in clear water. Absence of Jefferson rocks from the southwestern part of the region may relate to the Lemhi arch, a Middle to Late Devonian positive tectonic element in central Idaho, described by Sloss (1954). It is probably due to non-deposition rather than to post-Jefferson erosion, as suggested by the conformable contact of Three Forks shales and shaly limestones on the Jefferson near by.

The Three Forks was laid down over the entire region in a muddy-shelf sea during a period of relative tectonic stability, preceding a new cycle of tectonic differentiation. The sources of these sediments were a near-by lowland to the east, as shown by a regional increase in nonclastic Three Forks constituents. Deposition continued uninterruptedly into the Early Mississippian, as the Paleozoic geosyncline in the southwestern part of the region became more sharply differentiated from the tectonic shelf to the northeast. At first, the sea was still muddy at intervals as a result of periodic influxes of the clastic materials that now constitute the shale laminae between the limestones of the Lodgepole formation of the Madison group. Clearer seas prevailed at the end of Lodgepole time.

The clear sea persisted and probably reached its greatest extent in Mission Canyon time, when a thick sequence of relatively pure limestone and chert was laid down. In Middle Mississippian time an influx of muds from the geosynclinal region in central Idaho across the southwesternmost part of the Lima region took place, and Milligen shales accumulated on top of the Madison. Meanwhile, Madison limestones were still being laid down farther northeast. The eastward gradation of shales into limestones implies a westerly source for the clastics, which A. J. Eardley (Personal communication) believes may be found in a northerly extension of the Manhattan geanticline. Geosynclinal conditions persisted in the southwest during the remainder of Mississippian time, but the waters cleared once more, and thick limestones of the Brazer formation were deposited under conditions exceptionally favorable to the development of large cup corals. In the remainder of the Lima region the Madison sea may have lingered into Meramecian time; when finally it withdrew westward it exposed the surface of the Mission Canyon limestones to extensive leaching. During the Late Mississippian, while sedimentation of Milligen shales and Brazer limestones continued in the geosynclinal area, the sea encroached again on the shelf area to the north and east, receiving the shales and limestones of the Big Snowy (?) group. These shales may represent northeasterly tongues of the Milligen. Gypsum precipitated in temporarily isolated or semi-isolated basins.

Another brief southwestward retreat of the sea probably preceded deposition of the sandstones, shales, and impure limestones of the Amsden formation during latest Mississippian and earliest Pennsylvanian time. Gradually, the clastic material increased, and marine sands of the Quadrant formation accumulated across the entire region. The purity of these quartzites suggests a second-cycle derivation or a long period of weathering on the adjacent lowland to the east. The great thickness of the formation indicates a geosynclinal environment, or proximity to it. Quadrant deposition apparently continued through the Pennsylvanian Period, as evidenced by gradational contact with the overlying formation.

The Permian Period witnessed a change in the sea, resulting in accumulation of the limestones, shales, and cherts of the Phosphoria formation. Deposition was slow and took place under stagnant, reducing conditions, allowing concentration of hydrocarbons and phosphates in the shales. A temporary withdrawal of the sea may be indicated by a red bed near the base. Eastward gradation into a continental environment is suggested by common intercalation of red beds in the easternmost occurrences of the Phosphoria (McKelvey, 1949, p. 273). Thickness and phosphate content of the formation decrease in the same direction. Silica for the Phosphoria cherts may have been supplied by abundant quartzites exposed on this land, or by eruptions of the volcanoes that existed in central Idaho.

Mesozoic Events

The Mesozoic Era began with deposition of the calcareous shales and silts of the Dinwoody formation, probably continuous with the Phosphoria strata. The muddy Dinwoody sea provided an ideal environment for the burrowing brachiopod *Lingula*. Subsequent changes created a delta or flood plain on which accumulated the red clays of the Woodside formation, produced through intense chemical weathering on the adjacent landmass under humid-tropical or sub-tropical conditions. In late Early Triassic time the newly advanced Thaynes sea received influxes of mud from the lowlands to the west and north, where epeirogenic uplift took place (Eardley, 1951, pl. 11). This resulted in a series of calcareous shales and shaly and silty limestones. At the close of the Early Triassic the sea retreated, probably southward, and subaerial erosion cut a surface across the Triassic rocks, removing the Thaynes beds a short distance north of the Lima region.

No new invasion took place until Middle Jurassic (Bajocian) time, when the shallow Sundance sea spread across the Rocky Mountain region, and the Sawtooth, Rierdon, and Swift formations of the Ellis group accumulated in the southwestern part of the Lima region. The Sawtooth and Rierdon calcareous shales may have been partly derived from low positive

areas in the geosyncline to the west, but mostly from a southwesterly extension of the low "Belt Island" which continued into the northern part of the Lima region (Imlay *et al.*, 1948; Moritz, 1951, p. 1806–1807, Fig. 17; Schmitt, 1953, p. 367–371, 374, 379, Fig. 8). Oölitic Rierdon limestones were laid down in a shallow, agitated sea which had temporarily cleared. Moritz (1951, p. 1807) found that the insoluble residues of the limestones were derived from two provenances. A brief retreat of the sea preceded deposition of the glauconitic Swift sands, probably in part derived from the west or southwest (Schmitt, 1953, p. 387–388, Fig. 13), where pre-Nevadan uplifts occurred in the geosyncline.

Withdrawal of the sea at the close of the Oxfordian coincided with pre-Nevadan tectonic movements in other parts of the Cordilleran region. In Kimmeridgian time Morrison red beds accumulated on a lowland east of the pre-Nevadan uplifts. Local limestone beds probably were laid down in scattered temporary lakes. Swampy conditions prevailed at the end of Morrison time, forming black, carbonaceous layers, the equivalents of coal beds in other parts of the Rocky Mountains. The Nevadan orogeny to the west marked the end of Morrison deposition probably in early Portlandian time; nondeposition and, possibly, mild erosion occupied the remainder of the Jurassic.

From the beginning of the Cretaceous through the Barremian and Aptian stages, and probably as late as the Albian stage, large amounts of Kootenai clastics were worn down from the Cordilleran mountains and distributed eastward across this part of the Rocky Mountain region (Eardley, 1951, Pl. 15, p. 316). A conglomerate, possibly a pediment deposit, formed at the base, but, as the mountains were worn down, sandstones and shales accumulated in stream channels and on flood plains of large rivers. Repeated tectonic pulsations account for recurrent conglomerates higher in the section. In widespread temporary lakes on the flood plains limestones with abundant gastropod remains and a coquina bed at the top of the Kootenai were deposited.

Similar conditions continued through early Colorado (Aspen) time, with clay deposition greatly predominating over sands and conglomerates, indicating a more advanced stage of the erosion cycle. Explosive volcanic activity introduced large quantities of ash in the clays, which, upon alteration, became bentonites and porcellanites. The eastward regional decrease of volcanic material from the Aspen to the correlative Mowry formation reported by Dobrovolny (1940, p. 438) indicates that the volcanoes lay to the west, but the presence of volcanic breccia in the Aspen of the Lima area is evidence that at least some of the explosions took place not far away. The climate was warm-temperate, and the environment was well watered and vegetated, as shown by rather abundant plant fragments and by remains of a reptile adjusted to such a habitat (C. W. Hibbard, Personal communication).

Clastic deposition may have continued after Aspen time for an unknown period, but no post-Aspen sediments were recognized. Late Cretaceous erosion removed the entire Aspen formation from the central and northern portions of the Lima region.

Tertiary Events

Early Tertiary history of the Lima region was dominated by the Laramide orogeny. The first major phase produced broad, northeast-trending folds with steep southeast limbs, including the giant Blacktail-Snowcrest anticline and its two southwest prongs separated by the Little Water syncline (Fig. 3). Vigorous erosion bared the Precambrian core of the major uplift and produced the conglomerates and sands of the Beaverhead formation, which were spread across mountain slopes and along stream channels. Torrential drainage and landslides transported boulders up to 12 feet in diameter, one of which is exposed in the canyon of Indian Creek (West Fork), Idaho. A gradual increase in the intensity of uplift resulted in upward gradation into coarser clastics and eventually to tilting of the Beaverhead itself on the southeast flank of the major anticline.

The pre-Beltian and most of the Paleozoic pebbles in the conglomerate may have come from almost any direction but the only source of Belt and Kinnikinic pebbles was in the southwest. The coarser pebbles and cobbles must be of local derivation. In the southeastern

part deposition of finer conglomerates, sands, and silts prevailed. The probable Paleocene age of the Beaverhead sediments suggests that this first major disturbance was part of the mid-Laramide orogenic phase as recognized in Utah, Wyoming, and Idaho (Eardley, 1951, p. 242, 323, 331). Late Cretaceous (early Laramide) pulsations may have occurred, but apparently were not important.

Northeast folding was followed by new tectonic unrest, which produced northwest- to north-trending structures (Fig. 4). As this was introduced by an entirely new stress pattern, a considerable lapse of time probably separated the two disturbances, and the newer structures probably originated during the late Laramide (early Eocene) phase. This phase developed in several stages. The truncation of northwest folds by several thrust planes indicates that northwest folding marked the first stage. The new folds were superposed upon the earlier northeast structures and involved the Beaverhead conglomerates. This deformation was most intense in the western part of the region, where numerous anticlines were produced with generally steep to overturned easterly limbs. Farther east, Paleozoic and Mesozoic sediments had been stripped from large areas, and the crystalline core of the Blacktail-Snowcrest anticline apparently acted as a buttress against the new stresses. Northwest structures consequently are less prominent here.

Subsequent thrusting, in two steps, marked the second and third stages of the late Laramide orogenic phase. Both stages were restricted to the western half of the region, hence to the critical tectonic hinge zone between the edge of the pre-Tertiary tectonic shelf and the miogeosyncline to the west. First, the low-angle Medicine Lodge thrust and its probable branches, the Limekiln and Johnson thrusts, were produced at a depth of at least several thousand feet. The Medicine Lodge thrust sheet moved northeast possibly 10 miles or more, bringing a geosynclinal sedimentary rock facies close to a tectonic shelf facies. Meanwhile, continued folding and, locally, rather intense contortion took place. Finally, during the third stage of the late Laramide orogeny the high-angle Nicholia, Cabin, and Tendoy thrusts developed, but probably not until considerable sedimentary overburden had been eroded. The high-angle thrusts are believed to have broken through, dislocated, and tilted portions of the low-angle Medicine Lodge and associated thrusts. The thrust sheets overrode and, in part, incorporated elements of the earlier northeast and northwest folds.

The strains created when northeast compression was superposed athwart the earlier structures were locally resolved along high-angle normal and reverse faults. Another major event was the intrusion of a granite batholith in the Beaverhead Range, and of small igneous masses in the Snowcrest Range, probably during the latest stages of, or just after, the Laramide orogeny.

Following the Laramide orogeny, middle and late Eocene erosion reduced the region to probably late-mature relief. According to some, drainage on this surface was to the southeast (Umpleby, 1913a, p. 30; Atwood, 1916, p. 706–711; Shenon, 1928, p. 4). In late Eocene time basaltic lavas and clays, sandstones and conglomerates filled a narrow depression in the Sage Creek area, cut into the Beaverhead rocks. Drainage of this area must have been prevented by special, localized conditions, perhaps through damming by the lava flows or faulting. The fauna included several mammals adjusted to a subtropical lowland. Erosion continued uninterruptedly in other parts, and prior to the middle Oligocene exterior drainage was reestablished in the Sage Creek area, and the upper Eocene beds were locally eroded.

Extensive vulcanism broke out in middle Oligocene time. The Cook Ranch rhyolitic lavas and pyroclastics poured over much of the eastern half of the region. Similar materials in the Muddy Creek area farther west probably accumulated during this period. Local uplifts, perhaps the result of initial movements along the Muddy Creek and other high-angle faults, supplied coarse sands and conglomerates as well as finer clastics. These were washed into the adjacent valleys and became interbedded with the volcanic materials. Black shales accumulated in temporary lakes. The climate probably had become more temperate, but remained sufficiently warm and humid to permit the growth of Sequoia trees in forests inhabited by horses. Turtles and other sluggish animals

lived along the water courses. Fleet-footed animals ranged the open uplands.

Slight uplift and erosion of the Cook Ranch beds preceded early Miocene extrusion of the Blacktail Deer Creek basalts in the eastern half of the region. In the west, late Oligocene or early Miocene Medicine Lodge basalts and andesites and minor pyroclastics were extruded on the slopes of the Beaverhead Range, in the adjacent basin to the northeast, and at the edge of Muddy Creek Basin. Clastic materials, contributed by streams and mass wastage, accumulated on top of the Medicine Lodge volcanics and were interbedded with the Blacktail Deer Creek basalts. This period of sedimentation ended with deposition of freshwater limestone in extensive lakes bordering the Tendoy Range. Several major northwest-trending block faults (Fig. 5) may have helped to create the basins that received the lower Miocene sediments. Whether the Red Rock fault zone was active in creating the basin east of the Tendoy Range cannot be determined owing to the extensive Quaternary cover in this basin. The climate probably had become arid during the Miocene, suggested by the relatively large palatine pits in the oreodons from the Blacktail Deer Creek beds, a biological adjustment to a more arid climate in the opinion of Moojen (1948, p. 326–327, 393; Hibbard and Keenmon, 1950, p. 203).

By late Tertiary time low relief characterized much of the region. It truncated Tertiary basin beds and in places cut across the block faults to extend over the folded Beaverhead and pre-Beaverhead rocks of the present ranges. Monadnocks, whose summits preserved the earlier high surface, rose above it. Probably in Pliocene time rhyolitic lavas spread across this surface in the South Medicine Lodge area, and toward the close of the Pliocene or at the beginning of the Pleistocene high-level gravels were laid down in channels and on flood plains of major streams. Drainage on this surface was northeast and may have included the ancient Arco River. This drainage system was interrupted in latest Pliocene or early Pleistocene time by rapid elevation of the present mountain ranges, in part along major high-angle faults.

Quaternary Events

Uplift of the Beaverhead Range was too rapid for the Arco River to keep pace with it through erosion. Its upper course became separated from the lower course by the new Continental Divide along the crest of the Beaverhead Range, and reversed its flow. The lower course, now the Trail Creek-Nicholia Creek system, continued to the northeast, where movements along the northwest-trending Deadman, Kissick, and Red Rock faults were dislocating the late Tertiary erosion surface and lifting the present Tendoy Range. This uplift was slow enough for the stream and its tributaries to maintain their courses by cutting the deep gorge of Big Sheep Creek and the canyons of the lower Cabin and Deadman creeks. Farther east, the modern Blacktail Range probably rose along another block fault during this period. Tertiary beds in the adjacent basins were tilted.

This differential uplift of the ranges with respect to the basins may have been mechanically and chronologically related to the rise of a gentle east-west arch, complementary to the Snake River downwarp in central Idaho (Kirkham, 1927, p. 13, 24–26; 1931, p. 467–482). This arch presumably finds its topographic expression in the Monida and Bannock passes, two low divides crossing the basins on either side of the Tendoy Range. South of the axis of the arch the Edie School rhyolites and underlying basin bed limestones acquired a low regional southward dip, estimated by Kirkham at 100 to 125 feet per mile. Another event was the outflow of the Snake River basalts, either as separate extrusions on both sides of the southeastern end of the Tendoy Range or as tongues of the great flows of central Idaho. Basaltic extrusions near Armstead, in the Clark Canyon area, probably occurred simultaneously.

The downwarping of the Snake River region and the rise of the complementary arch to the north caused stream rejuvenation south of the arch axis. Deep canyons were cut into the freshwater basin limestones, rhyolites, and basalts by South Medicine Lodge Creek and its tributaries. No such canyons were formed in the

basins on the north side of the axis, where no comparable depression of the erosion base took place.

Meanwhile, the ranges were maturely dissected. In late Pleistocene (Wisconsin) time two successive periods of mountain glaciation occurred locally in the Beaverhead and Tendoy ranges. Several valleys developed cirques and U-shaped profiles, but otherwise were not deepened much by the glaciers, so that tributary valleys remained at grade. Bull Lake (?) and Pinedale (?) drift was deposited on the bottoms of these valleys and in the western portion of Big Sheep Creek Basin. Interglacial streams beveled the Bull Lake (?) moraine. In the nonglaciated basin areas terraces were formed.

In Recent time the basin streams gradually reached their modern levels, where they are now developing valley flats. The ranges were carved to their present mature topographies. Uplifts of the Tendoy Range along the Red Rock fault zone continue to this day. Large alluvial fans have been produced along the escarpment, and smaller fans are now building on top where the faults have been recently active. Occasional earth tremors probably are related to the Red Rock fault zone.

Economic Geology

Metallic Resources

Several small prospect pits for metals occur in the Lima region, but only one mine has been developed.

Sinkler (1942, p. 137) reports 2 per cent nickel and a trace of platinum in ultramafic sheets of the Blacktail Range, and small amounts of nickel in the Precambrian gabbros and ultramafics of the Ruby Range. Although locally of commercial tenor, the deposits are too small and remote to be of economic significance. Several gold-silver prospect pits occur south of the upper course of Meadow Creek in the Beaverhead Range, but no ore was observed. The prospects are in a green propylitic alteration zone in shattered brown basalt tuff and andesite. Similar epithermal mineralizations in Lemhi County, Idaho, described by Umpleby (1913a, p. 54–57) carry microscopic gold and silver. Small amounts of placer gold might be recoverable from the gravels along Horse Prairie Creek, west of Armstead, as some placer mining has been in operation along the tributary Jeff Davis and Colorado creeks, outside the northern limits of the Lima region (Trauerman and Waldron, 1940, Pl. I; Woodward and Luff, 1942, p. 403).

The Bonanza II lead-zinc mine south of Ellis Peak, in the western part of the Tendoy Range, has two levels at a 50-feet interval, in which open stoping has been practiced. About 14 carloads of ore were shipped from the railroad station at Armstead during 1948. The deposit, probably Tertiary, occurs in Mississippian limestones close to the fault contact with pre-Beltian rocks. The main vein is a tabular body striking N. 60° E. and dipping 40° S. Its maximum width is about 5 feet, and the vein has been followed longitudinally for approximately 175 feet. The main ore minerals are carbonates of lead and zinc. Small amounts of malachite occur as crustations in two quartz veins in the pre-Beltian complex east of Big Sheep Creek Basin. Chalcopyrite ore of noncommercial tenor occurs near the mouth of Jake Creek. Both types are probably Tertiary.

Finally, there is a possibility that the phosphatic "oil" shales of the Phosphoria formation contain sufficient concentrations of uranium and vanadium to be mined. Both metals are reported in many sedimentary phosphates (Clarke, 1924, p. 723–724; McKelvey, 1949, p. 270). Clarke also reports vanadium in "oil" shales. The combined resources of phosphate, hydrocarbons, uranium, and vanadium may prove to be of value.

Nonmetallic Resources

Phosphate.—The Phosphoria formation has yielded phosphate rock from several western States. As early as 1917 the Phosphoria occurrences near Lima and Dell were noted as potential sources of phosphate (Bowen, 1917, p. 315–320). Recently the formation has been investigated by the U. S. Geological Survey as part of a regional study of western phosphate deposits. The phosphate occurs disseminated in

two shaly units, referred to as units B and D by Klepper and Lowell (McKelvey, 1949, p. 273). McKelvey (1949, p. 279) reports that both shale units near Dell contain less than 32 per cent P_2O_5 and are therefore of no economic value at this time. Principal interest centers around the Snowcrest Range and the Centennial Range south of it, where the P_2O_5 content is 32 per cent.

The Phosphoria shales also contain small amounts of fluorine, vanadium, nickel, molybdenum, and uranium, in addition to solid hydrocarbons.

Gypsum.—In several localities concentrations of gypsum are present among the shaly and limy beds of the Big Snowy (?) group. One deposit in the Tendoy Range, north of the East Fork of Little Sheep Creek, was mined between 1940 and 1942, but considerable gypsum remains (Perry, 1949, p. 17). At this deposit the general northwest trend of the strata is complicated by a small cross syncline with steeply dipping limbs. Possibly, the gypsum was here concentrated by plastic flow during intense compression. These structural relations may afford a clue for future search for gypsum in this region. Perry estimates the original thickness of the gypsum bed at 10–15 feet.

Several other gypsum prospects are present in the Tendoy Range in the vicinity of Big Sheep Creek. Elsewhere, the gypsum bed may be absent or may be obscured by soil or talus.

Lime.—Several lime kilns were once in operation along North and South Limekiln canyons in the Tendoy Range but were abandoned many years ago. The Mission Canyon limestones of the Madison group were quarried and treated here.

Bentonite.—Beds of bentonite are intercalated in the Tertiary sedimentary deposits in several basins. Only locally are they of sufficient purity to be of industrial value. Bentonite of a fair degree of purity occurs also in the Aspen formation of the Sawmill Creek area, south of Lima. In almost all occurrences inaccessibility would make the cost of transportation prohibitive for commercial extraction.

Gravel.—Gravel deposits are abundant in the stream terraces of Red Rock Valley between Lima and Armstead. Locally, these beds are more than 20 feet thick. Several small pits along U. S. Highway 91 meet local highway and construction requirements.

Building stone.—The clean quartzitic Quadrant sandstones are a potential source of building stone. Other sources would be the Dillon granite gneiss and the Mississippian limestones. Most of the latter, however, are very cherty. Tertiary fresh-water and travertine limestones associated with various Tertiary basin sediments, could make a good decorative stone. The chief difficulty is, again, the inaccessibility of most of the outcrops. Those near railways and good roads may become of economic value for local supply.

Graphite.—The graphite deposits at the headwaters of Kate Creek have been described by Brink (1931, p. 13) and Perry (1948, p. 18–20). Prospecting was done around 1918, and many pits and adits were dug, but no graphite has been shipped. Although the average carbon content of 12 per cent compares favorably with other deposits, the Kate Creek graphite appears to be too fine-grained and to contain too much iron oxide. The graphite occurs in a zone about 1 mile long, 200 feet wide, and at least 200 feet deep along the fault contact between the pre-Beltian and Paleozoic systems. The amount of graphite varies considerably, and no dependable reserve estimate has been made. The graphite is believed to have been deposited in the fault fissure by hydrothermal solutions in post-Laramide time.

Coal.—Thin coal beds are intercalated in Tertiary basin beds at several localities. The coal is of mineable thickness along Keystone Gulch, a tributary to North Medicine Lodge Creek, where Peterson's coal mine is located. The mine has been operated intermittently, and the coal is sold to consumers at the mine. The product is a low-grade lignite. Analyses of samples from this locality are given by Dobbin (1932, p. 36).

Petroleum and natural gas.—The extreme southwestern corner of Montana has not generally been considered favorable for oil or gas (Condit, 1919, p. 22). The severe compression and intense fracturing and faulting in the Lima region, and the high uplift and deep erosion of a large part of the area, may have caused the destruction or escape of much of any oil and gas originally present. It is probable,

therefore, that the region does not have oil or gas potentialities of major proportions. On the other hand, the structural setting of the eastern part of the area is similar to that of several oil fields in western Wyoming and Alberta directly in front of large overthrust belts (Landes, 1951, figs. 149, 150; p. 515). Petroleum possibilities should, therefore, not be ruled out.

Favorable traps in the area may occur beneath the Tertiary basin beds. A well was drilled in Muddy Creek Basin around 1915, but was abandoned at 1000 feet without penetrating below the Tertiary beds (Winchester, 1923, p. 87). In 1953 a well (Cities Service, no. 1 Emerich) was drilled on the Lima anticline, W½ N.W.¼ N.E.¼ sec. 33, T. 14 S., R. 7 W., to a depth of 11,212 feet. It encountered steep dips along part of the way and was abandoned reportedly in the Morrison formation. The Lima anticline is a comparatively gentle structure superposed on the southeast flank of the Blacktail-Snowcrest anticline. Its southern limb is overridden by the Tendoy thrust. The trace of the Aspen-Beaverhead contact indicates a northwest pitch, but below this profound angular unconformity the regional southeast dip may offset the pitch, and the subsurface structure may plunge eastward. If so, any oil or gas originally present in the pre-Beaverhead rocks probably has escaped westward to the surface.

Potential petroleum reservoirs in the Lima region are the Kootenai and Quadrant sandstones and the cavernous top of the Madison limestone, which is productive in northern Montana. The porosity of the Quadrant quartzite, the equivalent of the highly productive Tensleep sandstone of Wyoming and central southern Montana, is extremely erratic at surface outcrops and may be largely related to ground-water leaching. Possible source rocks may include the Milligen and Big Snowy (?) black shales, but not the Phosphoria kerogen shales.

Oil may be obtained by distillation of the kerogen shales in the Phosphoria formation and the Muddy Creek basin beds. Both burn on a strong flame. Bowen (1917, p. 315–320) reports that the upper shale, which is about 100 feet thick, yields approximately 8 gallons of oil per ton of rock. Condit (1919, p. 24–26) reports figures between 16 and 26 gallons per ton for the Phosphoria shales at several places along Big and Little Sheep creeks. Commercial production of oil from these shales does not seem to be feasible at this time.

Ground water.—Little use is made of ground water in this vicinity. An increased utilization of this water for irrigation purposes would be beneficial to the hay crop and the cattle industry. Adequate supplies could be found in the buried stream gravels in the basins and in the large alluvial fans along the mountain fronts.

References Cited

Anderson, A. L., 1934, A preliminary report on recent block faulting in Idaho: Northwest Sci., v. 8, no. 2, p. 17–28.

—— 1947, Drainage diversion in the northern Rocky Mountains of East-Central Idaho: Jour. Geology, v. 55, p. 51–75.

Anderson, A. L., and Wagner, W. R., 1944, Lead-zinc-copper deposits of the Birch Creek district, Clark and Lemhi counties, Idaho: Idaho Bur. Mines Geology Pamph. 70.

Atwood, W. W., 1916, The physiographic conditions at Butte, Montana, and Bingham Canyon, Utah, when the copper in these districts was enriched: Econ. Geology, v. 11, p. 697–740.

—— 1917, Physiographic conditions and copper enrichment: Econ. Geology, v. 12, p. 545–547.

Blackwelder, Eliot, 1915, Post-Cretaceous history of the mountains of central western Wyoming: Jour. Geology, v. 23, p. 97–117, 193–217, 307–340.

—— 1917, Physiographic conditions and copper enrichment: Econ. Geology, v. 12, p. 541–545.

Bowen, C. F., 1917, Phosphatic oil shales near Dell and Dillon, Beaverhead County, Montana: U. S. Geol. Survey Bull. 661, p. 315–320.

Brink, L. D., 1931, Montana's graphite deposits: Mining Truth, v. 16, no. 2, p. 13.

Brown, R. W., 1933, Fossil plants from the Aspen shale of southwestern Wyoming: U. S. Natl. Mus. Proc., v. 82, art. 12, no. 2953.

Clarke, F. W., 1924, The data of geochemistry: U. S. Geol. Survey Bull. 770.

Cobban, W. A. and Reeside, J. B., Jr., 1951, Lower Cretaceous ammonites in Colorado, Wyoming and Montana: Am. Assoc. Petroleum Geologists Bull., v. 35, p. 1892–1893.

Condit, D. D., 1918, Relations of the late Paleozoic and early Mesozoic formations of southwestern Montana and adjacent parts of Wyoming: U. S. Geol. Survey Prof. Paper 120-F, p. 111–121.

—— 1919, Oil shale in western Montana, southeastern Idaho, and adjacent parts of Wyoming and Utah: U. S. Geol. Survey Bull. 711, p. 15–40.

Dobbin, C. E., *et al.*, 1932, Analyses of Montana coals: U. S. Bur. Mines Tech. Paper 529, 129 pages.

Dobrovolny, Ernest, 1940, Jurassic and Cretaceous

strata of the Camp Davis area, Wyoming: Mich. Acad. Sci. Papers, v. 26, p. 429–443.
Douglass, Earl, 1902, Fossil Mammalia of the White River beds of Montana: Am. Philos. Soc. Trans., n. ser., v. 20, p. 237–279.
—— 1903, New vertebrates from the Montana Tertiary: Carnegie Mus. Ann., v. 2, p. 145–199.
Eardley, A. J., 1947, Late Cenozoic trenches of the Rocky Mountains (Abstract): Geol. Soc. America Bull., v. 58, p. 1176.
—— 1951, Structural geology of North America: New York, Harper and Bros., 624 pages.
Emmons, W. H., and Calkins, F. C., 1913, Geology and ore deposits of the Phillipsburg quadrangle, Montana: U. S. Geol. Survey Prof. Paper 78, 271 pages.
Fenneman, N. M., 1930, Physical divisions of the United States (map): U. S. Geol. Survey.
Hausen, D. M., 1954, "Welded tuffs" of Oregon and Idaho: a hypothesis of origin (Abstract): Geol. Soc. America, Southeastern Section, Program of Meeting, p. 11–12.
Hayden, F. V., 1872, Preliminary report of the United States Geological Survey of Montana and portions of adjacent territories, being a fifth annual report of progress: Washington, D. C., 538 pages.
Heinrich, E. Wm., 1948, Pre-Beltian rocks near Dillon, Montana (Abstract): Geol. Soc. America Bull., v. 59, p. 1329.
—— 1949, Pegmatite mineral deposits in Montana: Mont. Bur. Mines Geology Mem. 28, 56 pages.
—— 1950, Sillimannite deposits of the Dillon region, Montana: Mont. Bur. Mines, Geology Mem. 30, 43 pages.
—— 1953, Pre-Beltian geologic history of Montana (Abstract): Geol. Soc. America Bull., v. 64, p. 1432.
Hibbard, C. W., and Keenmon, K. A., 1950, New evidence of the lower Miocene age of the Blacktail Deer Creek formation in Montana: Univ. Mich. Mus. Paleont. Contr., v. 8, no. 7, p. 193–204.
Honkala, F. S., 1949, Stratigraphy of the Centennial region, Beaverhead County, Montana (Abstract): Geol. Soc. America Bull., v. 60, p. 1897.
Imlay, R. W., Gardner, L. S., Rogers, C. P., Jr., and Hadley, H. D., 1948, Marine Jurassic formations of Montana: U. S. Geological Survey Oil and Gas Inv., Prelim. Chart 32.
Kirkham, V. R. D., 1927, A geologic reconnaissance of Clark and Jefferson and parts of Butte, Custer, Fremont, Lemhi and Madison counties, Idaho: Idaho Bur. Mines Geology Pamph. 19, 47 pages.
—— 1931, Snake River downwarp: Jour. Geology, v. 39, p. 456–482.
Klepper, M. R., 1951, A geologic reconnaissance of parts of Beaverhead and Madison counties, Montana: U. S. Geol. Survey Bull. 969–C, pp. 55–85.
Kummel, Bernhard, Jr., 1943, The Thaynes formation, Bear Lake Valley, Idaho: Am. Jour. Sci., v. 241, p. 316–322.
Landes, K. K., 1951, Petroleum Geology: New York, John Wiley and Sons, Inc., 660 pages.
LaRocque, A., and Edwards, C. D., 1954, Bear River and Aspen formations in Willow Creek, Teton County, Wyoming: Geol. Soc. America Bull., v. 65, p. 315–326.
Lowell, W. R., and Klepper, M. R., 1953, Beaverhead formation, a Laramide deposit in Beaverhead County, Montana: Geol. Soc. America Bull., v. 64, p. 235–244.
Mansfield, G. R., 1921, Types of Rocky Mountain structure in southeastern Idaho: Jour. Geology v. 29, p. 444–468.
—— 1924, Tertiary planation in Idaho: Jour. Geology, v. 32, p. 472–487.
—— 1927, Geography, geology and mineral resources of part of southeastern Idaho: U. S. Geol. Survey Prof. Paper 152, 453 pages.
McKelvey, V. E., 1949, Geological studies of the western phosphate field: Mining Eng. Trans., v. 1, p. 270–279.
Moojen, Joao, 1948, Speciation in the Brazilian spiny rats (genus *Proechimys*, family Echimyidae): Univ. Kansas Mus. Nat. Hist. Pub., v. 1, p. 301–406.
Moritz, C. A., 1951, Triassic and Jurassic stratigraphy of southwestern Montana: Am. Assoc. Petroleum Geologists Bull., v. 35, p. 1781–1814.
Newell, N. D., and Kummel, Bernhard, Jr., 1942, Lower eo-Triassic stratigraphy, western Wyoming and southeastern Idaho: Geol. Soc. America Bull., v. 53, p. 937–996.
Pardee, J. T., 1950, Late Cenozoic block-faulting in western Montana: Geol. Soc. America Bull., v. 61, p. 359–406.
Peale, A. C., 1896, Three Forks folio, Montana: U. S. Geol. Survey, Geologic Atlas of the United States, Folio no. 24, 5 pages.
Perry, E. S., 1934, Physiography and groundwater supply in the Big Hole Basin, Montana: Mont. Bur. Mines Geology Mem. 12, 18 pages.
—— 1948, Talc, graphite, vermiculite and asbestos in Montana: Mont. Bur. Mines Geology Mem. 27, 44 pages.
—— 1949, Gypsum, lime and limestone in Montana: Mont. Bur. Mines Geology Mem. 29, 45 pages.
Rich, J. L., 1918, An old erosion surface in Idaho: is it Eocene?: Econ. Geology, v. 13, p. 120–136.
Richards, R. W., and Mansfield, G. R., 1912, The Bannock overthrust, a major fault in southeastern Idaho and northeastern Utah: Jour. Geology, v. 20, p. 681–709.
Ross, C. P., 1934, Correlation and interpretation of Paleozoic stratigraphy in south central Idaho: Geol. Soc. America Bull., v. 45, p. 937–1000.
—— 1937, Geology and ore deposits of the Bayhorse region, Custer County, Idaho: U. S. Geol. Survey Bull. 877, 161 pages.
—— 1947, Geology of the Borah Peak quadrangle, Idaho: Geol. Soc. America Bull., v. 58, p. 1085–1160.
Schmitt, G. T., 1953, Marine Jurassic in Rocky Mountains: Am. Assoc. Petroleum Geologists Bull., v. 37, p. 355–393.
Schuchert, Charles, and Dunbar, C. O., 1941, Outlines of historical geology: New York, John Wiley and Sons, Inc., 291 pages.
Scott, H. B., 1935, Some Carboniferous stratigraphy in Montana and northwestern Wyoming: Jour. Geology, v. 43, p. 1011–1032.
Shenon, P. J., 1928, Geology and ore deposits of the

Birch Creek district, Idaho: Idaho Bur. Mines Geology Pamph. 27, 125 pages.
Shimer, H. W., and Shrock, R. R., 1944, Index fossils of North America: New York, John Wiley and Sons, Inc., 837 pages.
Sinkler, H. M., 1942, Geology and ore deposits of the Dillon prospect, southwestern Montana: Econ. Geology, v. 37, p. 136–152.
Sloss, L. L., 1946, Fauna of the Amsden and Heath formations: Mont. Bur. Mines Geology Mem. 24, p. 6–11.
—— 1950, Paleozoic sedimentation in Montana area: Am. Assoc. Petroleum Geologists Bull., v. 34, p. 423–451.
—— 1954, Lemhi arch, a mid-Paleozoic positive element in south-central Idaho: Geol. Soc. America Bull., v. 65, p. 365–368.
Sloss, L. L., and Moritz, C. A., 1951, Paleozoic stratigraphy of southwestern Montana: Am. Assoc. Petroleum Geologists Bull., v. 35, p. 2135–2169.
Stearns, H. T., Crandall, Lynn, and Steward, W. G., 1938, Geology and water resources of the Snake River Plain in southeastern Idaho: U. S. Geol. Survey Water Supply Paper 774, 268 pages.
Stearns, H. T., Bryan, L. L. and Crandall, Lynn, 1939, Geology and water resources of the Mud Lake region, Idaho: U. S. Geol. Survey Water Supply Paper 818, 125 pages.
Tansley, Wildred, Schafer, P. A., and Hart, L. H., 1933, A geological reconnaissance of the Tobacco Root Mountains, Madison County, Montana: Mont. Bur. Mines Geology Mem. 9, pt. 1, 57 pages.
Trauerman, C. J., and Waldron, C. R., 1940, Directory of Montana mining properties: Mont. Bur. Mines Geology Mem. 20, 135 pages.
Umpleby, J. B., 1912, An old erosion surface in Idaho: its age and value as a datum plane: Jour. Geology, v. 20, p. 139–147.
—— 1913a, Geology and ore deposits of Lemhi County, Idaho: U. S. Geol. Survey Bull. 528, 182 pages.
—— 1913b, The old erosion surface in Idaho: Jour. Geology, v. 21, p. 224–231.
Weller, J. M., *et al.*, 1948, Correlations of the Mississippian formations of North America: Geol. Soc. America Bull., v. 59, p. 91–196.
Wilmarth, M. G., 1938, Lexicon of geologic names of the United States: U. S. Geol. Survey Bull. 896, 2 vols., 1244 and 1151 pages.
Winchell, A. N., 1914, Mining districts of the Dillon quadrangle, Montana, and adjacent areas: U. S. Geol. Survey Bull. 729, 191 pages.
Winchester, D. E., 1923, Oil shale of the Rocky Mountain region: U. S. Geol. Survey Bull. 729, 202 pages.
Wood, H. E., 1934, Revision of the Hyrachidae: Am. Mus. Nat. History Bull. v. 67, art. 5, p. 181–295.
—— 1941, Nomenclature and correlation of the North America continental Tertiary: Geol. Soc. America Bull., v. 52, p. 1–48.
Woodward, G. E., and Luff, Paul, 1942, Gold, silver, copper, lead and zinc in Montana: U. S. Bur. Mines Minerals Yearbook, p. 389–413.
Yen, Teng-Chien, 1952, Age of the Bear River formation, Wyoming: Geol. Soc. America Bull., v. 63, p. 757–763.

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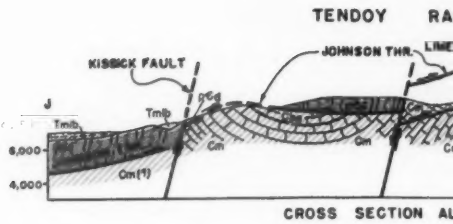
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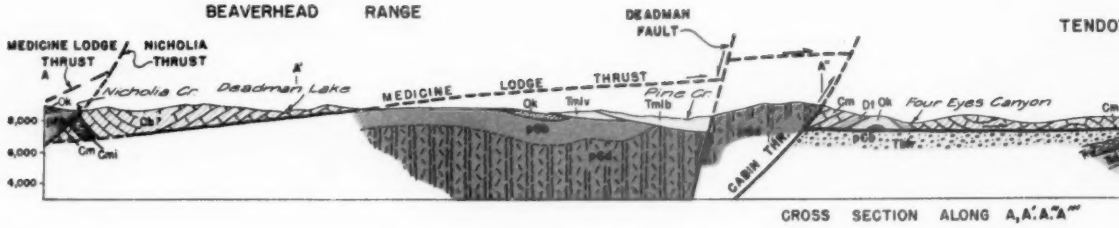
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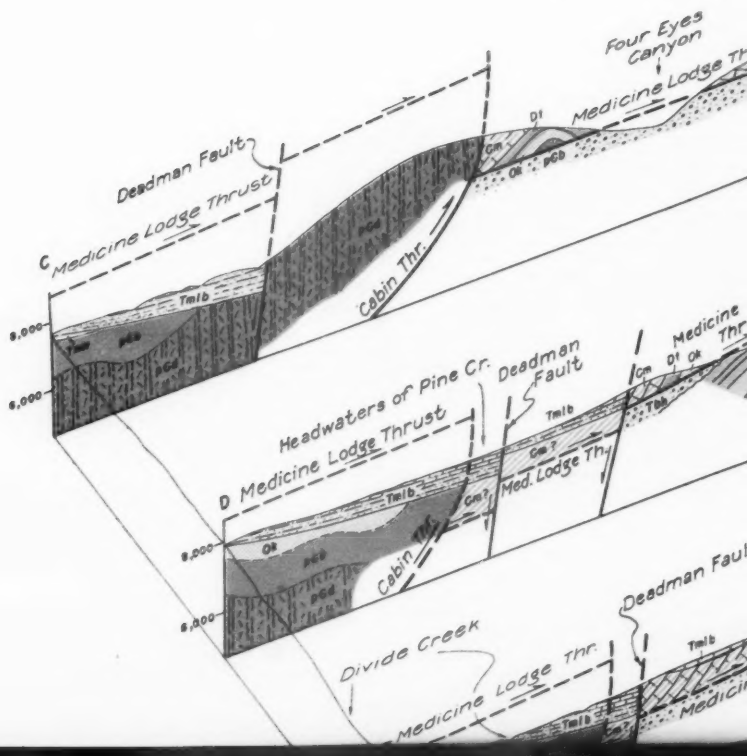
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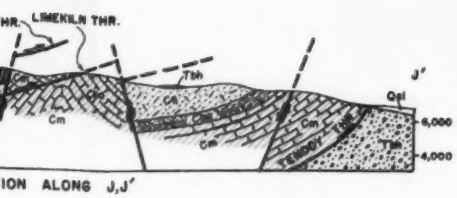

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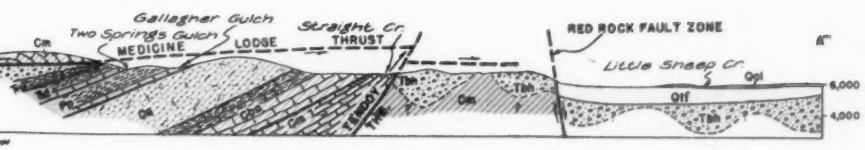

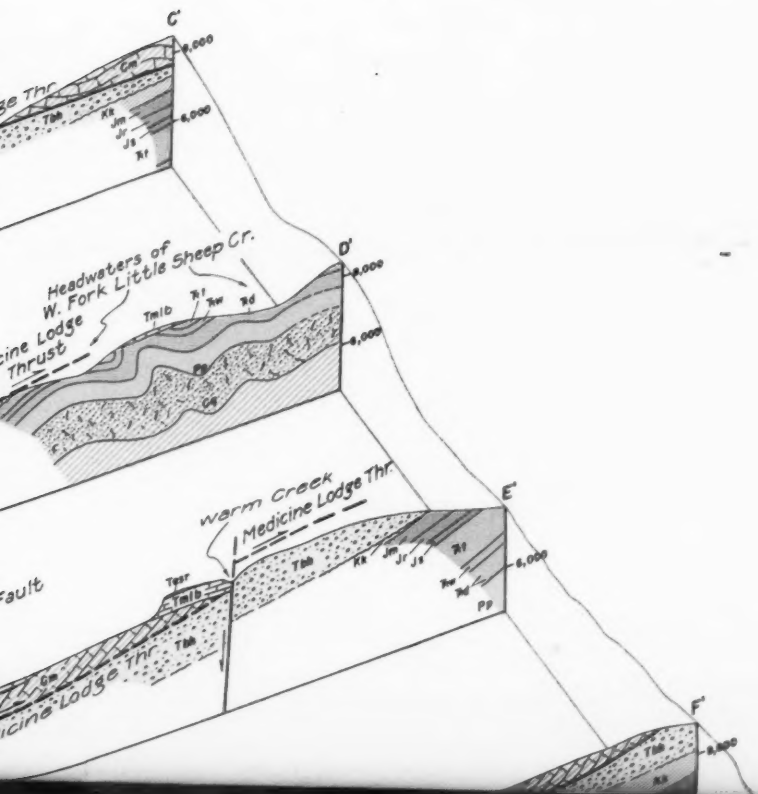

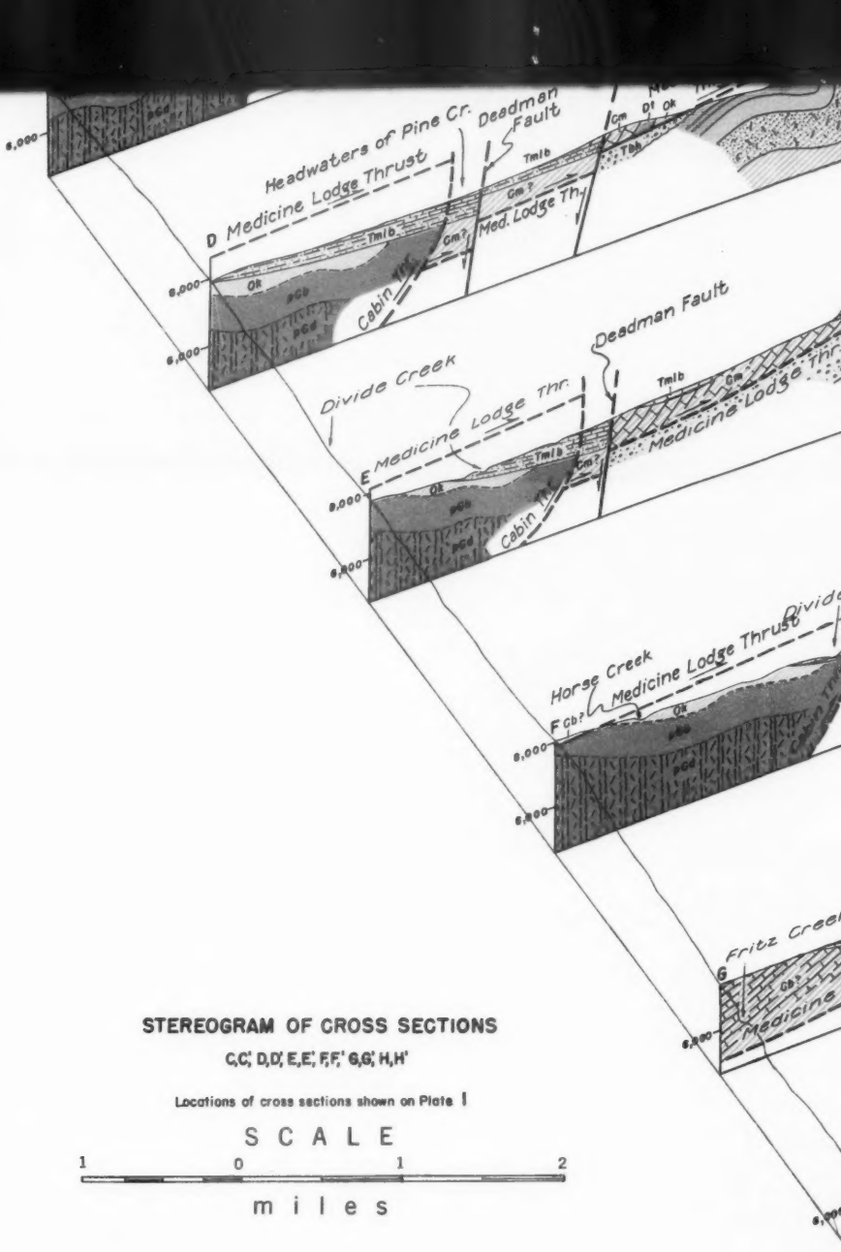

STEREOGRAM OF CROSS SECTIONS

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Locations of cross sections shown on Plate I

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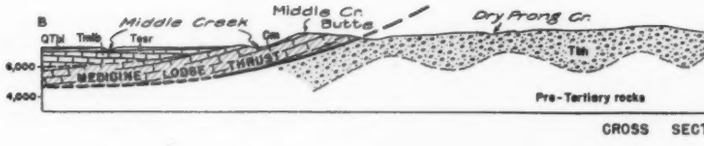

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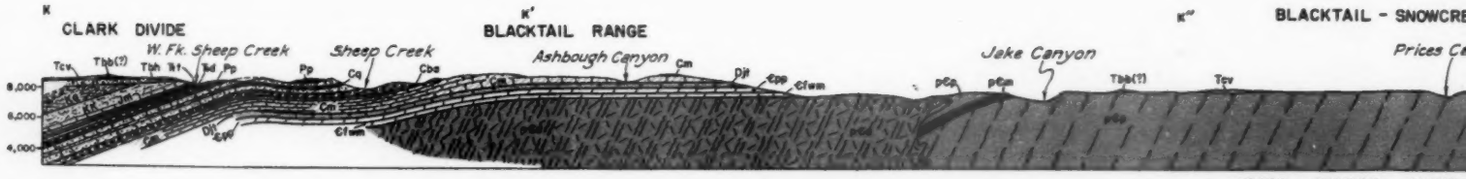

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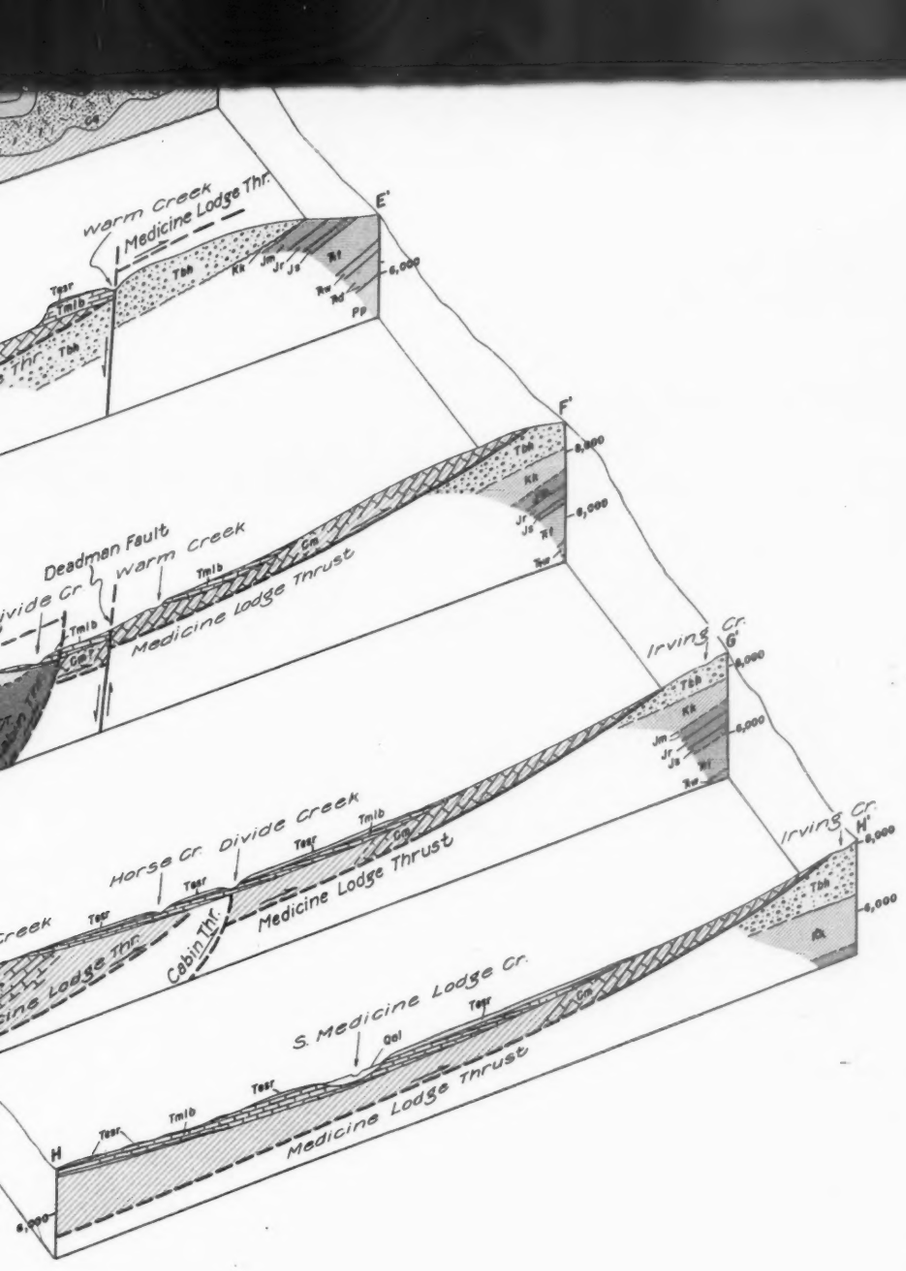

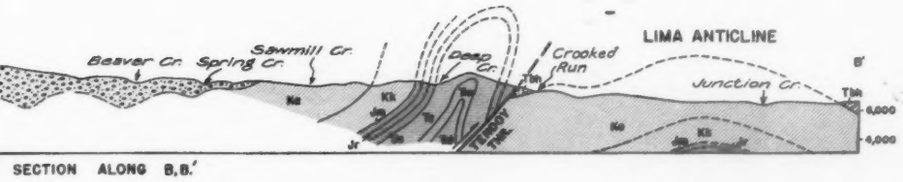

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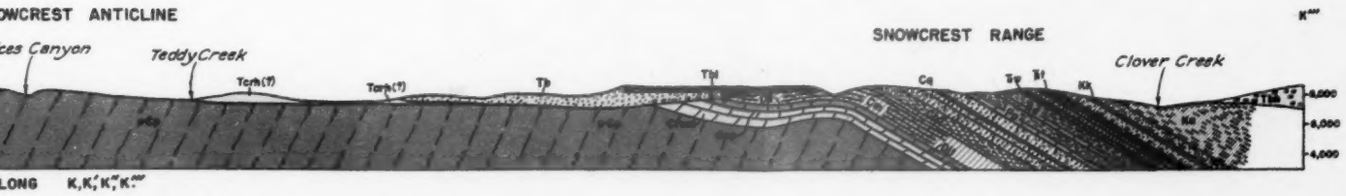

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DISTRIBUTION OF IGNEOUS ROCKS IN SPACE AND TIME

By G. W. Tyrrell

Abstract

Consanguinity in an assemblage of rocks, associated in time and place, is held to indicate that it has been derived by a more or less uniform geological process from a common initial magma. The most fundamental grouping of igneous rocks is thus based on distributional relations, which are in turn dependent on geological process and event.

The close association of igneous activity with earth movement is well known. The repetition of petrographic "provinces" or "kindreds" in time and place is clearly a manifestation of the cyclic character of tectonic events in geological history. The tectono-igneous cycle is elaborated and discussed with special reference to Scotland and Northwest Europe, and comparisons are made with eastern North America and Indonesia.

CONTENTS

TEXT

	Page
Introduction	405
Consanguinity	406
Kindreds of igneous rocks	407
Petrographical provinces and periods	408
Igneous action and tectonics	409
Geological cycle	409
Diastrophic subdivision of the crust	409
Diastrophic (tectonic) cycle	410
Igneous cycle, with special reference to Scotland, Northwest Europe, and eastern North America	412
Quartz dolerite dikes and sills of North Britain	415
Tectono-igneous cycle in Indonesia	416
Plutonic and volcanic associations	417
Nature and origin of igneous cycles	420
References cited	424

TABLES

Table	Page
1. Correlation of the igneous and tectonic phases of the Tertiary and Quaternary of Indonesia with those of the Paleozoic of Scotland	418
2. Tectono-igneous cycle	420
3. Comparison of the igneous phases of the Caledonian, Hercynian, and Alpine orogenic cycles	421

Introduction

My assignment today is to introduce the subject of your deliberations—*The distribution of igneous rocks in space and time,*—a subject which, while it concerns high-temperature materials, will probably not generate so much heat as the symposium 4 years ago introduced by my friend Professor H. H. Read. It is, however, a subject which has attracted much attention in the past 60 years and has given rise to a voluminous literature. Nevertheless it has not yet advanced far beyond the stage of fact collection, and the important generalizations which I feel are inherent in this field on the relations of igneous activity to tectonics and its correlation with the geological cycle are still largely to be made.

In recent years many of what we used to think were characteristic igneous rocks have been relegated to the category of metamorphism (plutonism in Read's sense) and with, I think, some justification. With some reservations I am disposed to accept Read's (1948, p. 3) classification of rocks as *neptunic* (mainly marine sediments), *volcanic* (magmatic, igneous, dominantly effusive and basic in composition), and *plutonic* (metamorphic, migmatitic and granitic, of deep-seated origin). The last-named group also includes, beside granite, many formerly described as plutonic igneous rocks.

Because of the doubts about the origin of many plutonic rocks I am now disposed to rest generalizations on the distribution of igneous rocks in space and time mainly on the undoubted igneous rocks which have clearly been derived from liquid or quasi-liquid magmas,

together with the intrusive rocks, sometimes of quite coarse textures ("plutonic"), which are associated with them—that is, on what W. Q. Kennedy (Kennedy and Anderson, 1938, p. 25) has called the volcanic association. But I still think that consideration of many of those types now assigned to Read's plutonic group can help us toward the above-desired end.

By a curious coincidence some groups of lavas have a very close chemical resemblance to plutonic groups, as rhyolites with granites; dacites, latites, and andesites with members of the granodiorite - tonalite - quartz - diorite group; trachytes with alkali syenites; phonolites with nepheline syenites; and basalts with gabbros. I cannot help thinking that in at least some cases there is a genetic connection between lavas and the corresponding plutonic rocks. And this leads me to question the complete divorce in geology, petrology, and genesis of the volcanic from the plutonic rocks advocated by Kennedy (Kennedy and Anderson, 1938, p. 31) and Read (1948, pp. 3–4). It is true that both authors concede that certain intrusive dolerites, gabbros, and other "plutonic" rocks may sometimes be included in volcanic associations.

Discussion as to what are and what are not igneous rocks has been hampered by an unnecessarily restricted definition of magma as "completely fluid rock substance" (Read, 1948, p. 5), and F. F. Grout (1948, p. 46) has rightly animadverted upon it. If this definition is accepted literally, practically the only truly magmatic rocks are the pure glasses like obsidian. When examined after sudden quenching even the basaltic liquids of the lava lakes (*e.g.*, Halemaumau) are seen to have had a few crystals in them, and there are plenty of instances of both effusive and intrusive action in which the magmas concerned were rich in crystals. The writer knows of instances of the Lower Carboniferous plagioclase basalts (Markle type) of the Midland Valley of Scotland in which the feldspar crystals exceed 50 per cent by volume of the rocks, and are sometimes so clotted together as to simulate an anorthosite, but which yet form lavas and small intrusions. It would be more accurate to say that a magma is a mush of liquid and crystals that is capable of effective intrusion or extrusion,—a material something like porridge in which crystals correspond to the oat grains. This view is confirmed from the etymological point of view by my colleague Professor A. W. Gomme[1], to whom I applied for information as to what the term *magma* meant to the ancient Greeks.

It is no answer to say that salt, clay, and certain metamorphic rocks must therefore be regarded as magmatic because, under some circumstances, they seem to conform to the above definition. The writer maintains that rock-flowage materials can readily be distinguished from magmatic when the associated phenomena are taken into consideration. A perfect transition must exist, and many Transformists are willing to admit it, between just-mobilized granitic material and granitic magma which might consolidate as a granitic vein or as a flow of rhyolite; but we are not called upon to fix an arbitrary boundary between mobilized material and magmatic material any more than we are bound to fix a point at which white ends and black begins in the perfect transition of shades of gray which intervenes between white and black. Common sense applied to the observation of relevant facts must decide such issues, and there is room for different opinions.

Consanguinity

It is a generally accepted fact that certain groups of igneous rocks show various degrees of family likeness (Judd) or consanguinity (Iddings). This community of character is expressed in their chemical, mineral, textural, and geological features. While consanguineous suites may range in chemical composition from acid to ultrabasic, may exhibit many kinds of textures, and may consist of extrusions or intrusions or both, some mineral and chemical characters, which may be described as *constant*, are common to practically all members. Other characters are *serial* and show a regular variation throughout the group which may be illustrated by suitable graphs. Thus, in some

[1] Professor of Greek, University of Glasgow. Professor Gomme says the word comes from a root meaning to *knead*, especially to knead in a mould as, for example, a kind of barley cake which was not baked but eaten in the soft state. The past tense (kneaded) was also used of a poultice. The term *magma* itself is actually found only in some Greek medical literature for a thick unguent; and Webster's New International Dictionary, 1st. Ed. (1932 issue, p. 1295) confirms that it is still used in this sense, but only rarely, by pharmacologists.

suites, a constant chemical character is oversaturation with silica; highly silicated feldspars and quartz appear in small quantity even in some of the basic members. Regular variations of silica, alkalis, ferrous iron, magnesia, and lime may serve as examples of serial characters. My late friend and colleague, M. A. Peacock, indeed founded a very successful criterion for the discrimination and comparison of igneous series on the "lime index"—a relation between silica, alkalis, and lime. Some suites or series may be characterized throughout by a peculiar mineralogical feature, such as the occurrence of anorthoclase in Norwegian, East African, and Antarctic suites, or the occurrence of orthorhombic pyroxenes in even the acid members of the so-called charnockite series. Further, whatever views we may hold regarding the origin of micropegmatite in certain dolerite and gabbro sills and dikes, its presence together with some characteristic geological features in great basic suites of world-wide distribution may be cited as yet another criterion of consanguinity.

Consanguinity in an assemblage of igneous rocks, together with spatial and temporal propinquity of its members, may be held to indicate that it has been derived by a more or less uniform geological process from a common initial magma, or from a number of closely related magmas, however such magma or magmas have been generated.

Kindreds of Igneous Rocks

The suggestion the writer made in 1926 (Tyrrell, 1926a, p. 136) that, "Groups, series or suites of igneous rocks which show consanguineous chemical and mineral characters, and which appear to be genetically related", should be called *kindreds* has not been taken up to any extent by petrographers, but it has value in relation to the subject of the distribution of igneous rocks. The term kindred in igneous petrology has somewhat the same meaning as the term *facies* in stratigraphy and sedimentary petrology. Several terms have been used in much the same sense as, for example, series, suite, tribe, clan, branch, and stem (*stamm*), but none appears so suitable as kindred[2] for the major groups of igneous rocks. The relationship implied by consanguinity may be of different degrees of closeness; the term kindred is here used for the widest and most tenuous kind of family likeness as that, for example, between the members of the alkalic or calcic kindreds. The writer further suggested that the progress of research might reveal that kindreds included *tribes*, the rocks of which would show a closer relationship than those of kindreds; and that tribes might be found to be composed of *clans*, the members of which would show the highest degree of consanguinity of all. Both the alkalic and calcic kindreds, if indeed they are true kindreds, probably contain several tribes and clans.

By their nature kindreds, tribes, and clans cut across the more formal groupings and classifications, and certain abundant rock types such as basalt, andesite, and "granite" (if I may use this term to denote an igneous rock) may recur in several different groupings. A kindred will be composed of a wide range of igneous rock types, but all or many are marked by common chemical and mineral characters which may be exhibited in constant or serial relations. Other factors may serve to establish the individuality of kindreds. They may be characterized by the different bulk development of their component types. Thus, in one kindred, andesites greatly predominate in the range basalt—pyroxene andesite—hornblende-mica-andesite—dacite—rhyolite; while another consists almost entirely of basalts (*e.g.*, basalt floods).

Futhermore, there are differences among kindreds in regard to the relative bulk of "plutonic," hypabyssal, and volcanic developments within the accessible crust. While this feature may, of course, be due to accidents of geological environment, erosion, and exposure, the writer believes that there are some intrinsic differences in this respect. Finally, kindreds may be distinguished by their extension in space and time (petrographic provinces and periods) and by their varying relations to tectonic processes. These topics will be further developed in later sections of this paper. In *Principles of petrology* (1926a, p. 137–141) the

[2] The terms kindred, tribe, and clan are based on analogy with those used for peoples and languages. For example, the Aryan kindred comprises (or is supposed to comprise) tribes (Celtic, Hellenic, Teutonic, etc.), and the tribes include clans with a closer degree of relationship than obtains within the other two categories. The terms kindred, tribe, and clan are therefore here used in the proper hierarchical order.

writer describes and discusses certain kindreds regarded as having been tentatively established, but it is unnecessary to go furtehr into that subject here. The systematization of kindreds, tribes, and clans has barely begun. In all probability the necessary data are at hand in the vast accumulations of petrographical literature which have piled up during the last 30 or 40 years, ready for the master mind and pen of a petrological Suess.

Petrographical Provinces and Periods

In the writer's opinion the most fundamental grouping of igneous rocks, as for all rocks, is in relation to geological process and event. In effect this means a grouping based on distributional relations. The idea of petrographical provinces was dimly foreshadowed by several writers in the middle of the nineteenth century, but, as Lowell says, "the credit for an idea belongs to him who utters it clearest and best", and the credit for this idea therefore belongs to J. W. Judd. Judd's definition (1886, p. 54) is:

"There are distinct petrographical provinces within which the rocks erupted *during any particular geological period* present certain well-marked peculiarities in mineral composition and microscopical structure, serving at once to distinguish them from the rocks belonging to the same general group, which were simultaneously erupted in other petrographical provinces".

A. Harker (1911, p. 3), with the superior literary ability which distinguished his writings, improved on Judd's definition with the following:

"A petrographical province is a more or less clearly defined tract within which the igneous rocks *belonging to a given period of igneous activity* present a certain community of petrographical character traceable throughout all their diversity, or at least obscured only in some of the extreme members of the assemblage."

In both these definitions the phrase italicized (by the present writer) insists on a time limitation. In other words a petrographical province has extension in time as well as in space. Many objections to the idea have been based on the mistaken notion that the term petrographic province applied to *all* the igneous rocks within a given region regardless of their ages.

The writer's definition, crystallizing the idea in terms of kindreds, is that

"A *petrographical province* is the geographical extension of a kindred, and a *petrographical period* is likewise its extension in time" (Tyrrell, 1926a, p. 141).

A petrographical province therefore, like existence itself, is a space-time concept. The boundaries of petrographical provinces and periods are necessarily vague and ill-defined because the true spatial and temporal associations of a kindred are with a particular geological environment and tectonic process which also have ill-defined boundaries in space and time. It is therefore better to approach the concept of petrographical provinces by placing the emphasis more on the nature of the kindred itself than on its mere geographical or temporal extension. For this reason the geographical names which have been applied alike to petrographic provinces and to the assemblages called kindreds, such as Atlantic for alkalic kindreds, Pacific for calc-alkalic kindreds, Arctic (Von Wolff, 1914) for the great basalt floods (surely a most unsuitable term!), and Mediterranean (Niggli, 1921) for certain assemblages characterized by richness in potash, are to be deprecated as useless and misleading.

The very wide and vague but, the writer believes, valid grouping of igneous rocks into *alkalic* and *calcic* has been much exploited in the past in connection with petrographic provinces. It is highly probable that both terms cover several distinct kindreds, and that some kindreds pass over the gradational boundary between the two groups, thus being "mixed" in the sense that they contain both alkaline and calcic types. But, despite Whitman Cross's animadversion on this point (1912), this very character is one of those which distinguish certain kindreds as, for example, those of the Scoto-Irish Tertiary region and the Hawaiian volcanic field. Harker (1911, p. 5) greatly overemphasized the importance of the alkalic-calcic contrast in using such phrases as: "A given province is either of calcic or alkalic facies typical members of the two branches not being found together"; and again: "It is possible to map out the active parts of the earth's crust into great continuous regions of alkalic rocks on the one hand and of calcic rocks on the other". This perhaps merited Cross's severe strictures (1912, p. 366–367); but in justice to Harker it must be pointed out that he further said (p. 4):

"This two-fold division of igneous rocks is, of course, in no wise a final or exhaustive treatment of

the subject; but as a first step towards a natural or genetic classification it seems to be established beyond question".

W. Q. Kennedy (Kennedy and Anderson, 1938, p. 25–29) has divided the igneous rocks into plutonic and volcanic associations and has related them to orogenic and nonorogenic regions of the earth's crust respectively. This concept is clearly relevant to the ideas of kindreds and petrographic provinces. He also pointed out that, as the orogenic zones in time become welded to the adjacent stable regions of the earth's crust, and then participate in their epeirogenic reactions, it is natural for different kindreds to be successively established within the same region. Thus the view that petrographic provinces should have time limitations is confirmed from another angle. The writer defers further reference to Kennedy's views until discussion of the tectonic aspect of the problem.

Igneous Action and Tectonics

That igneous activity is closely associated with earth movement is one of the fundamental observations of geological science, but the nature and laws of the connection are still only little understood notwithstanding recent contributions to the subject by Kennedy and Anderson (1938), A. Holmes (1932), H. H. Hess (1938), W. H. Bucher (1933), A. Rittmann (1936), and many others. For further advances we need to know much more of kindreds and petrographic provinces and of their correlations with earth movements. Closer study of these involved questions must still await more exact discrimination of types and sequences of earth movements and of their significance in the economy and history of the earth's crust. Even more needed is a deeper understanding of the igneous rocks (in a double sense!), of the criteria for their natural grouping, and of the chemical composition and variations of kindreds, tribes, and clans. The close association of igneous activity and earth movement in space and time makes it natural to assume a causal connection between them or, better, that they are concomitant effects of the same causes operating in and beneath the crust. The work of A. Holmes (1928) on subcrustal currents induced by radioactive heat in relation to tectonirs and igneous activity constitutes a penetrating exposition of one of these causes, although one must not forget its hypothetical nature.

Geological Cycle

The concept of the geological cycle expresses the idea that geological events fall into more or less regular sequences which may be repeated at more or less regular intervals; but in the writer's view the geological cycle is compounded of five parallel and correlated subcycles of tectonic (or diastrophic), igneous, metamorphic (or plutonic), sedimentational, and erosional events, and it is the integration of all into a connected whole that constitutes *the* geological cycle.

Nature, like history, never repeats herself—exactly! Or, to put it in the more usual way, Nature, like history, repeats herself—with variations! While geologists recognize that the events of geological history repeat themselves again and again in a more or less regular sequence or rhythm, they do not expect this rhythm to be exactly repeated or invariable. In other words the geological them has variations imposed upon it. A fairly exact analogy is the musical Theme with Variations. Thus Brahms, in his "Variations on a Theme by Haydn", takes that composer's noble tune, states it simply at first, and follows with a number of variations in which, however, the attentive ear can always discern the underlying theme. The analogy breaks down, of course (as all analogies eventually must), when the composer turns the tune completely upside down or inside out, as he occasionally does! The geological cycle is a complex conception consisting, as it does, of several interwoven subcycles, but it can be resolved in comparatively simple terms, especially if the component subcycles are described separately.

Diastrophic Subdivision of the Crust

The term diastrophism covers the sum total of earth movements; and, for the purposes of this paper, diastrophism can be roughly divided into epeirogeny and orogeny. Vertical movements in the oceanic sectors have been termed *thallatogenic* by Kober (1921), but the writer regards these as coming under the heading of epeirogenic. No attempt can be made here to distill the essence of the vast recent literature on diastrophism and tectonics; only that which

is considered relevant to igneous action is selected.

The incidence of diastrophism varies both in time and space. In a given region of the crust orogeny is episodic and may entirely cease for long periods. Epeirogeny, however, is long-continuing and probably never entirely ceases. There seems to be a diastrophic rhythm; epeirogenic movement rises to a maximum and falls to a minimum. Maximum orogenic movement and igneous action appear to coincide in time with the epeirogenic crescendo. Periods of maximum diastrophism have been termed *orocratic* by W. Ramsay (1924). Orocratic periods are relatively short and are separated by *pediocratic* periods in which epeirogenic movement sinks to a minimum and orogenic movement often does not occur at all. Joly estimated the general ratio of duration between orocratic and pediocratic periods at 1:3.

It has long been recognized that there are strong, rigid, stable portions of the earth's crust (shields) that have not suffered folding, or any diastrophism except gentle vertical movements, since very early geological time. At the same time there exist certain narrow, elongated, subsiding areas (geosynclines) in which, or near which, thick sedimentation followed by orogeny takes place at intervals. At any given period, according to Kober (1921), the earth's crust may be divided into *kratogens* (cratons in American literature), strong rigid blocks some of which approximate continental dimensions, separated from one another by *orogens*, narrow, elongated, weak, labile zones whereon, in the fullness of time, fold mountains arise. This contrast is fundamental to our subject as the igneous, sedimentary, and other cycles are widely different in orogens and kratogens.

S. von Bubnoff (1923, p. 4), while adhering to Kober's contrast of kratogens and orogens, attempts a more detailed tectonic classification of crustal elements, two of which are stable and two variable:

(1) Permanent continental blocks with a steadily rising tendency

(2) Permanent oceanic sectors with a steadily subsiding tendency

(3) The slightly mobile continental shelves, varying constantly between shallow seas and lowlands

(4) The highly mobile orogens alternating between comparatively deep subsidences (geosynclines) and high fold mountains.

He holds that the diastrophic relations of these four crustal elements are fundamentally different and find expression in their diverse magmatic and sedimentational associations.

While kratogens such as the continental nuclei remain approximately in constant position (*pace* Wegener!), the sites of the orogens which develop around them shift from time to time. Old orogens become tectonically "dead" and are so welded onto the adjacent kratogens as to form integral parts of them, acquiring kratogenic characters and reacting in a kratogenic manner to subcrustal causes of movement in succeeding geological cycles. Northwestern Europe affords an excellent example of this phenomenon for, around the Fenno-Scandian Shield, the loci of orogenic activity in successive cycles since the Precambrian (Caledonian, Hercynian, Alpine) have shifted gradually from the northwest to the southeast and south, and each orogen, in its turn, has been welded onto the circum-Baltic kratogen.

The above and later remarks are amplified from the writer's incomplete paper of 1937 presented to the Seventeenth International Geological Congress at Moscow which was published in English in Volume I of the Report of the Seventeenth Session dated 1939, but which only reached the western geological world in 1948.

Diastrophic (Tectonic) Cycle

A very simple conception of the diastrophic cycle, sufficient for much of our present purpose, is that it begins with the breakdown of the broad high continents ridged with fold mountains, which represent the climax of the previous cycle. The earth's crust begins to sag off the edges of the continents into geosynclines which are weighted with immense loads of sediments derived from the erosion of the adjacent highlands. At the same time the continents become lower and smaller, the oceans broader and shallower, while the geosynclines continue to deepen until some equilibrium is attained between the operative and opposing forces. Toward the end of the geosynclinal period the oceans, filled with sediments, become very broad and shallow, spilling over the lowland continental margins in great trans-

gressions. The continents, now small and low, contribute little but fine argillaceous sediments to the seas, wherein the main sedimentation is organic (limestones, etc.). Meanwhile tectonic forces have been gathering strength beneath the seeming quiet and, toward the end of the cycle, break out, with perhaps one or two preliminary spasms, in revolutionary power, buckling the geosynclinal sediments into folds which become lifted above sea level, first as island arcs, and then as great fold-mountain ranges. Hence the diastrophic cycle may be held to consist of:

(1) A geosynclinal or evolutionary phase, occupying perhaps three-quarters of the cycle, and

(2) An orogenic or revolutionary phase, much shorter and comprising both orogenic and epeirogenic movements

As an aftermath of the orogenic phase, both outside and within the orogenic arcs, the earth's crust may break into large blocks which suffer vertical movements along deep-reaching major faults. This usually begins during the waning stages of the orogenic phase or somewhat later and ushers in a new diastrophic cycle.

Many igneous cycles fit in well with this simple version of the tectonic cycle. A more elaborate scheme is that by R. A. Sonder (1922, p. 242–247); based mainly on tectonic events in Northwestern Europe, it fits the writer's particular field of study very well. Sonder divides the diastrophic cycle into four phases[3]:

(1) *Geosynclinal* phase, in which epeirogenic movements become strong, leading to alternations of continental uplift and oceanic subsidence. Geosynclines come into existence near the continental margins which, as they deepen, are kept filled by sediments eroded from the adjacent highlands. There is a resurgence of igneous activity of ophiolitic type within the geosynclines. Toward the end of this phase there may be small preliminary orogenic spasms bringing island arcs to the surface over the geosynclinal areas, and accompanied by volcanoes yielding andesitic and rhyolitic lavas.

(2) *Transgression* (Fluctuation) phase. The continents are now worn down to minimum size and height, and climate has become more or less equable over the whole earth. The oceans have transgressed widely over the edges of the continents and are broad and relatively shallow. Rivers bring down only small amounts of muddy sediments to be deposited as shales in warm shallow seas swarming with life, where they are augmented by limestones due to the accumulation of remains of the abundant marine organisms. Epeirogeny, orogeny, and igneous activity are minimal or absent, but toward the end of the phase diastrophic unrest begins again, to culminate finally in the succeeding phase.

(3) *Orogenic* (Emergence) phase, beginning with regression of the seas from the old continents, associated with intense compression, folding, and mountain building along the former geosynclinal zones. This is accompanied and followed by vertical movements giving rise to broad continents surmounted by fold-mountain ranges, with great volcanic activity at the surface, and plutonism (metamorphism, granitization, and magmatic granite) in the depths of the orogenic zones.

(4) *Continental* phase, during which orogeny is in abeyance, the continents attain their maximum size and remain comparatively stable. There is still some vertical movement and considerable volcanic activity. Extremes of climate result in the formation of great ice sheets and extensive deserts with the deposition of glacial and "red-rock" sediments.

The use of Sonder's scheme in this connection does not imply its full acceptance, or that it is applicable all over the earth. Von Bubnoff (1924, p. 173) has criticized it on the ground that, far from being a generalization from world geology, it is really based on the sequence of geological events in Northwestern Europe and eastern North America, both highly complicated and disturbed regions in which the sequence of events may not be the normal one. Further, it does not take into account the fact that the earth's crust consists of dissimilar units, with different histories, and most probably with differing and nonsynchronous geological cycles. Moreover, geological cycles do not always coincide with the standard geological eras. The great Russian Platform, for example, has remained almost undisturbed since Precambrian time, and must have experienced a quite different cycle of events from those of geosynclinal and orogenic regions such as western Europe and eastern North America. Despite

[3] Sonder begins his cycle with the Continental Phase. The writer prefers to begin it with the Geosynclinal Phase for reasons which will appear later.

these criticisms Sonder's cycle does seem to fit events in at least the Paleozoic of northwestern Europe, and seems therefore best adapted as the frame within which the sequence of igneous events can be discussed.

Igneous Cycle, with Special Reference to Scotland, Northwest Europe, and Eastern North America

The Paleozoic succession of igneous rocks in Scotland affords an excellent example of an igneous cycle parallel to the simplified tectonic cycle (Tyrrell, 1926, p. 284). It may be represented as follows:

Geosynclinal Phase	Early:	*Early Ordovician* Ophiolitic series; basalts, spilites, keratophyres, etc. Albite dolerite, gabbro, serpentine, etc.
	Late:	*Late Ordovician-Middle Silurian* Little or no igneous activity
Orogenic Phase	Early:	*Late Silurian?* Foliated granites and granodiorites, augen gneiss, oligoclase gneiss, *lit-par-lit* gneiss, etc. in concordant sheet intrusions and injection complexes (?). Metamorphosed ultrabasic and basic igneous rocks; foliated lamprophyric sills and dikes.
	Late:	*Late Silurian-Lower Old Red Sandstone* Intrusions of serpentine, gabbro-, and norite. Granite, granodiorite, tonalite, and quartz diorite in massive discordant batholiths, sheet- and ring-intrusions; associated with numerous small bodies of kentallenite, appinite, hornblendite, scyelite, etc. Andesitic and rhyolitic lavas.
Post-Orogenic Phase		*Carboniferous-Permian* Basaltic lava fields with olivine basalt, trachybasalt, trachyandesite, trachyte, and phonolite. Associated intrusions, mostly sills, of picrite-teschenite, theralite, essexite[4], monchiquite. Quartz dolerite sills and dikes.

[4] Luscladite-Lacroix

Sonder's diastrophic cycle is peculiarly suitable, for reasons given above, as a frame for discussion of the Paleozoic igneous succession in Scotland, supplemented by reference to the corresponding Scandinavian and North American igneous fields. The use of eastern North America in this connection, although it lies 2000 miles or more west of the Scottish and Scandinavian scene, is justified by the probable consideration that the North Atlantic did not exist in Paleozoic time (Bailey, 1928, p. 76); or better, perhaps, that North America was much closer to Northwest Europe until at least the end of the Paleozoic. The above fields present evidence of one complete diastrophic and igneous cycle (in Sonder's sense) together with the closing phase of an earlier one, from the Late Precambrian to the Permian.

(4A) Continental phase. Late Precambrian (Torridonian) and Early Cambrian

(1B) Geosynclinal phase. Late Cambrian and Early Ordovician (Scotland)

Igneous rocks belonging to this phase are not represented in Scotland, but the equivalent Scandinavian formations (Dala, Jotnian, Sparagmite, etc.) contain flows of amygdaloidal basalt, with great sills of dolerite (mainly quartz dolerite but some olivine dolerite) and a dike swarm consisting of the same rocks. The equivalent North American formation (Keweenawan) contains the enormous Lake Superior basalt flood of estimated volume 24,000 cubic miles (C. R. Van Hise and C. K. Leith, 1911), which is probably associated with extensive dike swarms of dolerite and two of the largest gabbro lopoliths (Duluth and Sudbury) in the world.

Submarine volcanic activity along the geosynclinal zones gave rise to an ophiolitic (or spilitic) kindred. The volcanic rocks may be normal basalts, but there are many spilitic and some keratophyric types relatively rich in soda as compared with potash, and with albite or oligoclase as the dominant feldspar. Pillow structure is often spectacularly displayed as an indication of subaqueous eruption. The accompanying hypabyssal and (old-style) plutonic rocks are sills and dikes of albite dolerite, and masses of gabbro, pyroxenite, and peridotite (serpentine)—an ophiolitic assemblage (Steinmann, 1926, p. 637). These rocks are

frequently involved with the later orogeny and altered to the so-called "green rocks" or greenstones (chlorite-, hornblende-, glaucophane-, talc schists, "prasinites", etc.). The constant submarine eruptions and the accompanying earthquakes attested by fine slumping phenomena (Henderson, 1935) sent clouds of basic debris mingled with the more normal sediments of the sea floor up through the water. The gradual settlement of this material led to the formation of the characteristic graded graywackes and shales which are now recognized by many writers as the typical sedimentary rocks of the geosynclinal phase along with cherts (often radiolarian) and chemical limestones (Tyrrell, 1933; Krynine, 1941; Pettijohn, 1943; Krumbein and Sloss, 1951). Dewey and Flett (1911, p. 246) and Steinmann (1905; 1926, p. 662) had already associated the spilitic (ophiolitic) suite with slow oceanic subsidence; and Von Bubnoff (1923, p. 65) had pointed out the association of these rocks with geosynclinal depression.

Ophiolitic zones of the same Early Paleozoic ages extend southwestward from the Southern Uplands of Scotland into the center of Ireland and northeastward, along the entire length of the Scandinavian peninsula (Goldschmidt, 1916, p. 7–21), right up into Spitsbergen, a total distance of some 3000 miles. The same rocks are no doubt found on the same scale in eastern North America (*e.g.*, in Newfoundland); but the writer has neither time, space, nor competence to discuss the American equivalents as thoroughly as they deserve. This task must be left to American geologists who are more conversant with the terrain.

(2B) Transgression (Fluctuation) Phase. Late Ordovician and the greater part of the Silurian in Scotland

In this phase there was only slight epeirogenic movement, and, in Scotland, igneous activity was wholly in abeyance.

(3B) Orogenic (Emergence) Phase. The Caledonian Orogeny. Late Silurian and Early Devonian (Lower Old Red Sandstone)

In Scotland there seem to have been at least two peaks of igneous activity. The earlier was the injection of foliated granites and granodiorites, oligoclase gneiss, etc., in comparatively small masses as concordant intrusions before and during the folding (synorogenic). In Sutherland there is a group of ultrabasic and basic intrusions and augen gneisses, associated with Moinian and Lewisian rocks, which have been highly metamorphosed along with these formations. In addition there are two great injection complexes (injection gneisses, permeation gneisses, granite gneisses, granites, and the peculiar intermediate and basic hybrid rocks known as the Ach'uaine type), which Read (1931, p. 11–12) assigns to a pre-Torridonian date. Also, in the Grampian Highlands, associated with rocks similar to the above-mentioned acid gneisses, is a group of greenstones (amphibolite, epidiorite, hornblende schist) and large bodies of ultrabasic rocks (now serpentine and associated types), which were intruded prior to the metamorphism of the Dalradian (Read, 1948, p. 39). Their exact date therefore depends on which is considered to be the age of the Dalradian. It is now all but certain that Cambrian rocks are involved with the Dalradian (J. G. C. Anderson, 1947, p. 482–484), and therefore the most probable date of the "Older Igneous Rocks" and the Dalradian metamorphism is Post-Cambrian—*i.e.*, they belong to the earliest part of the Caledonian orogeny (Late Silurian). D. L. Reynolds (1942, p. 43) also regards the "Older Granites" as migmatites belonging to a syntectonic episode of the Caledonian orogeny, although she might demur to calling them igneous. The writer therefore regards all these rocks, with the possible exception of the Sutherland complexes, as belonging to the earliest phase of Caledonian orogeny. They probably represent the products of plutonic processes (in Read's sense) which took place at the roots of the geosynclinal prisms.

The assemblage is unevenly distributed throughout the Scottish Highlands, and rocks of the same types and presumably of the same age extend up the length of the Scandinavian peninsula from south to north, especially in the western (Norwegian) part. From a very cursory dip into the literature the writer believes that the igneous rocks of both parts of the Orogenic Phase are not so widely distributed in eastern North America as they are in Northwest Europe, but he is open to correction on this point by American geologists more familiar with the ground.

The second period of igneous activity occurred after the folding had ceased and presumably when deep-seated plutonic processes had reached their end stage of complete mobilization, with the production of eruptible magma capable of truly igneous injection into the upper parts of the orogenic zones. There seem to be two distinct series. One, which has apparently contributed many boulders of similar rocks to the conglomerates of the Downtonian and Lower Old Red Sandstone, is most probably of Late Silurian age. It consists of a group of ultrabasic and basic sheet intrusions (peridotite, olivine gabbro, troctolite, olivine norite, hypersthene gabbro, etc.) occurring in Aberdeenshire and Banffshire; and a group consisting of massive intrusions of biotite granite, granodiorite, tonalite, and diorite with a hypabyssal entourage of porphyrites, lamprophyres, aplites, and pegmatities, which are distributed throughout the Central Highlands and include the great Cairngorm mass and the Aberdeen granites. They are all massive, unfoliated, and structureless (except for lineations and what one may call the Cloos *Gefuge*). In their relations to the country rocks they are mostly crosscutting and discordant, though with some suggestions of flattish tops and even bases (Read, 1948, p. 48–51).

The second series is in part later than the Lower Old Red Sandstone and is best developed in the Southwest Highlands. It consists of plutonic masses of granite, granodiorite, tonalite, and quartz diorite (Ben Nevis, Glen Etive, Ben Cruachan, etc.), associated with dike- and sill-swarms of porphyries and lamprophyres. An abundant suite of small bosses of monzonite, kentallenite, and cortlandtite, some of which are relatively rich in potash, are associated with the above acidic types. Some of these basic rocks also occur in intrusive complexes along with granite and granodiorite. It is in this series that the remarkable cauldron ring complexes of Ben Nevis, Glen Coe, and Glen Etive are developed (Read, 1948, p. 53–57).

A group of large and small plutonic masses of granite, granodiorite, and quartz diorite associated with a small number of basic types and accompanied by dike swarms of lamprophyre and porphyry, occurs in Galloway and the Southern Uplands of Scotland south of the Midland graben. It is not yet known whether these masses are to be associated with the first or second of the post-orogenic groups, but certain large intrusions of felsite along or near the Southern Boundary Fault seem to be intrusive into the Lower Old Red Sandstone.

The above post-orogenic suite was preceded and accompanied by andesitic and rhyolitic eruptions from small volcanoes, forming the Lorne, Sidlaw, Ochil, Pentland, Carrick, and Cheviot Hills. The principal lava types erupted were highly feldspathic "basalts" with sporadic olivine[b], pyroxene-, hornblende-, and biotite andesites, dacites, and rhyolites. At Taynuilt (Argyllshire) in the Lorne field there are relatively potash-rich lavas called orthoclase-olivine basalt which may correspond to the kentallenite type among the "plutonic" rocks (Bailey and Maufe, 1916, p. 185). These lavas were erupted onto a land surface as there are sedimentary intercalations and cavity fillings of Old Red Sandstone type, marked with the tracks and trails of crawling animals. Many of the lavas have reddened tops and rusty patches which may mark the sites of temporary pools (Tyrrell, 1914, p. 66–68).

A great unconformity exists between the Lower and Upper divisions of the Old Red Sandstone during which period the new continent reached stability. In one or two parts of the Midland Valley of Scotland a final spasm of Caledonian orogeny caused gentle open folding in the Lower Old Red Sandstone. (*Cf.* connection of andesitic volcanoes with this type of folding.)

(4B) Continental Phase. Latest Old Red Sandstone, Carboniferous and early Permian

The tectonic history of the continental phase begins with the breakdown of the Old Red Sandstone continent by block faulting, with the formation of a great graben or rift valley in the Scottish midlands along northeast to southwest faults—*i.e.*, in the main Caledonian direction. The Midland Graben is bounded to the north by the Highland Boundary Fault which, as it affects the Lower but not the Upper di-

[b] The writer prefers to call these rocks *olivine andesites*. In chemical composition they are very different from the true olivine basalts of the Carboniferous.

vision of the Old Red Sandstone, must have been formed in Middle Old Red Sandstone times. To the south the graben is bounded by an echelon of faults which may be collectively referred to as the Southern Uplands Boundary Fault. Both these fault systems run in the general northeast to southwest Caledonian direction. The rift depression is filled with Old Red Sandstone and Carboniferous rocks; and its floor is broken by differential fault movements into a mosaic of smaller blocks. This faulting continued right through the Carboniferous and after it. As time went on the faults veered more and more to the east and west direction, thus becoming parallel to the major direction of the Hercynian folds which were coming into existence some 400 miles to the south in Southern England.

Igneous activity in the Midland Valley of Scotland is associated with the initiation and early history of the Midland Graben. It began in late Upper Old Red Sandstone time with the eruption of mainly olivine basalts from hundreds of small volcanoes of central type, giving rise to comparatively large basalt fields (Tyrrell, 1931a, p. 186–188). The major period of eruption was early Carboniferous, but the activity continued with ever-diminishing strength through the epochs of the Carboniferous Limestone Series and the Millstone Grit, and flared up finally in the Carbo-Permian. All this igneous activity seems to be definitely earlier than the east-west faulting. The lavas belong to an olivine basalt differentiation series leading to trachyte and phonolite (trachybasaltic kindred—A. Holmes) at least in the greater part of the petrographic period; but the later eruptions (Late Carboniferous-Early Permian) are coarse olivine-rich basalts (Millstone Grit) and analcite- and nepheline-bearing olivine basalts (Early Permian). Neither group is accompanied by trachytes or phonolites. Their hypabyssal representatives are differentiated sills of teschenite-picrite-peridotite, bekinkinite, theralite, and monchiquite, showing fine examples of gravity differentiation, and forming what is probably the largest petrographic region of analcite-bearing igneous rocks known. Dikes associated with this phase of igneous activity are scarce except around some large centers of trachytic eruption. No "plutonic" equivalents are known at the present level of erosion, except fragments of peridotite, pyroxenite, anorthosite, gabbro, etc. in some vent agglomerates[6].

In Scandinavia the probable equivalent of this Scottish period-province is the famous alkalic province of Oslo, Norway, now known to be Permian. In eastern North America it is the White Mountain Magma Series of New England (together with the Monteregian field of the adjacent part of Canada?), of which the age is now believed to be Mississippian (Williams and Billings, 1938, p. 1025). Possibly the New Hampshire Magma Series, syntectonic with the Acadian Revolution, is the tectonoigneous equivalent in New England of the Scottish granodiorite-andesite kindred which is roughly syntectonic with the Caledonian Revolution (Billings and Keevil, 1946, p. 810–811). The writer invites the attention of American geologists, especially those whose field of work is in the New England States, to the close similarity among the Oslo, Scottish, and New England petrographic provinces in point of age, petrographic composition, and igneous structures. Such a comparison may suggest another line of evidence which would lead us to suspect that North America was much nearer Northwest Europe in the later Paleozoic than it is now.

Quartz Dolerite Dikes and Sills of North Britain

To complete the consideration of the Paleozoic igneous cycle in Scotland let us turn now to discuss a late igneous event quite different from those dealt with above—namely, the intrusion of a great suite of quartz dolerite dikes and sills (F. Walker, 1935; J. E. Richey, 1939; A. G. MacGregor, 1948). They occur as thick (60–100 foot) extended dikes which run in a general east-west direction, some of them for scores of miles. They are associated with a number of thick sills which, although only one direct connection has been proved, have cer-

[6] In one of these vents, that of the Black Rock at Fairlie (Ayrshire), the writer recently found a block of carbonatite rich in soda hornblende, closely resembling types from Fen (Norway) and Alnö (Sweden). In others the writer has found biotites like those described by H. S. Washington from Italy. Xenocrysts of anorthoclase, titanaugite, soda hornblende, and biotite are common.

tainly been fed from the dikes. They are probably slightly earlier than the picrite-teschenite sills and are thus Permo-Carboniferous. Their area of distribution is much more extensive than that of the preceding basic alkalic suite, which is largely confined to the Midland Graben. They cover an area from Aberdeenshire on the north to south of the English Border, where the Great Whin Sill of the North of England appears to be the southernmost and largest member of the suite.

The chemistry and mineralogy, the mode of differentiation or, rather, the almost complete lack of differentiation, the prominence of thick dikes, and the much wider area of distribution of this suite are in complete contrast to the preceding alkalic suite. The magmatic composition of these rocks and their geological occurrence, so alien to those of the Carbo-Permian, support the view that they are derived from new and different magmatic sources, and that they may mark a completely new beginning in the sequence of igneous events. It can also be surmised that the quartz dolerite magma was derived from more deeply seated sources than the limited and localized reservoirs of the Carbo-Permian igneous rocks which, as we have seen, diminished greatly in bulk and became highly differentiated and specialized toward the end of their eruption period. The alkalic magmas thus seem to have been almost completely exhausted before the arrival of the quartz dolerites.

In the writer's opinion the east to west extension of the quartz dolerite dikes and the limitation of their age to the earliest Permian give a clue to their tectonic affiliation. The writer believes they represent a phase of igneous activity in the northern kratogen corresponding to the Hercynian orogeny which had been proceeding in the geosynclines far to the south. A. G. MacGregor (1948, p. 148) relates them to the Borcovician stress system which operated from late Carboniferous to earliest Permian time. The situation is rather involved and difficult to disentangle. There is here perhaps what may be called a sudden transfer of allegiance from one tectono-igneous system to another. The alkalic rocks of the Carbo-Permian graben were clearly associated with the waning Caledonian orogeny, and with block faulting of the Old Red Sandstone continent as compression relaxed after the cessation of folding. The alkalic magmas, erupted from local and comparatively shallow reservoirs, were gradually exhausted, and the province became magmatically "dead". The Scottish region then became part of a great kratogen bounded to the south by the rising Hercynian folds. The tensions thus created caused fissures which extended deep enough to tap the intermediate layer of olivine-free basalt (Kennedy and Anderson, 1938, p. 35–39; MacGregor, 1948 p. 148).

The above, of course, raises difficult questions of the contemporaneity of tectono-igneous cycles proceeding in different, although adjacent, parts of the earth's crust, and of variations in such cycles. At the same time (Late Silurian to Lower Devonian) that an orogenic phase was in progress in the Caledonian orogen, a geosynclinal phase accompanied by an ophiolitic series (Devonian) was proceeding in the Hercynian orogen. The question is also raised as to whether we ought to treat orogens and kratogens separately in regard to tectono-igneous cycles. The writer thinks we should. These questions are far-reaching in their implications, and it is doubtful whether the data are yet at hand for their solution. Their further exploration must be left for future work.

Tectono-Igneous Cycle in Indonesia

For an example of a more recent tectono-igneous cycle we may turn to Indonesia which has been closely investigated by R. W. van Bemmelen (1949; 1950)[7]. He starts from a correlation between petrographic provinces and the phases of structural evolution (diastrophic cycle) from the Cretaceous to the present time within the great Tertiary orogen of that region. Starting from the south an "Atlantic suite" of igneous rocks occurs in or near the southern foreland of the orogen (Christmas I. etc.), followed by an "ophiolitic suite" in the adjacent foredeep to the north which is of Late Cretaceous to Paleogene age. Then come three successive orogenic phases in the course of the Tertiary and Quaternary, each accompanied by the metasomatic emplacement of "acid

[7] The writer has only very recently come across van Bemmelen's book on the *Geology of Indonesia* (1949), and has not yet had time fully to understand and assimilate his views on igneous origins.

plutonites" (quartz diorite, granodiorite, granite) in the cores of the geanticlines. Each orogenic phase was succeeded by a cycle of "basalto-andesitic" volcanicity in the Oligo-Miocene, Mio-Pliocene, and Quaternary, respectively, the lavas being erupted from the crests of the geanticlines. In at least the second orogenic phase the "rising front of granitization" invades the lavas belonging to the first cycle of orogenic volcanism ("Old Andesite Formation"). The first and second of these cycles were terminated by enormous eruptions of "acid volcanites" (mainly tuffs)[8] respectively generated by and in connection with the upward advance of the acid plutonites of the second and third orogenic phases. Toward the end of the third orogenic phase in Java some aberrant differentiates of "Mediterranean affinities" were produced besides the normal "Pacific" andesitic lavas (*e.g.*, the Muriah volcano with potassic lavas). The post-orogenic igneous rocks of the cycle consist of large spreads of "tholeiitic plateau basalts" which cover the northern hinterland of the Indonesian orogenic arcs.

Van Bemmelen's summary of the above sequence is given in the first three columns of Table 1, to which the writer has added two columns showing what he could gather of the petrographic characters and ages of the igneous rocks of the Indonesian cycle, and the corresponding stages in the Scottish region. The comparison makes clear a parallelism between the two sequences. From Van Bemmelen's rather compressed and involved style of English writing it is sometimes difficult to extract the relevant facts of age and petrographic composition, and there are a few apparent inconsistencies in his account. His Indonesian cycle starts with a Pre-orogenic phase which the writer is inclined to regard as the final continental phase of the preceding cycle. There is no clear indication of a Transgression phase. Within his Orogenic phase there appear to be three tectono-igneous subcycles, and there are also indications of three in the Scottish field—namely, the pre- and synorogenic phase at the beginning of the Caledonian orogeny, and two post-folding phases, respectively Late Silurian and Lower Old Red Sandstone. There is even a phase of potash-rich magmas in both the Indonesian and Scottish fields at approximately the same position in the sequence of tectono-igneous events. The Indonesian cycle concludes with a Post-Orogenic Phase of fissure eruptions emitting tholeiitic basalt lavas which find their parallel in the Scottish quartz dolerite dikes and sills; but there seems to be no indication of an Indonesian trachybasaltic kindred belonging to the Post-Orogenic Phase. The general parallelism between the Indonesian and Scottish tectono-igneous cycles thus appears to be very close.

[8] In a summary account of van Bemmelen's 1950 paper (*Science Progress*, July 1951, p. 506) the writer inadvertently wrote "plutonites" instead of "volcanites".

Plutonic and Volcanic Associations

In an important paper W. Q. Kennedy (Kennedy and Anderson, 1938) put forward the fruitful conception of plutonic and volcanic associations. The plutonic association includes all subjacent igneous masses (concordant and discordant batholiths, stocks, sheet complexes, etc.) which are intruded at the roots of an orogenic belt during or immediately after an active orogeny, together with their associated minor intrusions, but with practically no effusive equivalents. The volcanic association, on the other hand, consists mainly of superficial effusive rocks, with the intrusive masses genetically associated with them and derived from the same magmatic sources, within a limited cycle of volcanic activity. The intrusions of the plutonic association are overwhelmingly acid, consisting almost entirely of granodioritic and granitic types. Typical basic and ultrabasic rocks are characteristically rare or absent. The volcanic association, on the other hand, is composed mainly of basalts and their less abundant differentiates which belong to the two basaltic lines of descent postulated by Kennedy. Acid lavas and intrusions are thus rare in comparison with the basic. True plutonic associations are connected with orogenic belts; but, although Kennedy recognizes volcanic associations within the orogenic zones he is unwilling to regard them as genetically connected with the plutonic rocks, the great bulk of volcanic effusions and intrusions occur within, and are genetically connected with, the stable, non-

TABLE 1.—CORRELATION OF THE IGNEOUS AND TECTONIC PHASES OF THE TERTIARY AND QUATERNARY OF INDONESIA WITH THOSE OF THE PALEOZOIC OF SCOTLAND

Petrographic provinces	Orogenic zones	Zonal stages of orogenic evolution	Age and petrographic characters	Scottish Paleozoic equivalents
I. Atlantic Suite	Foreland (1)	Pre-Orogenic	Trachybasaltic kindred of Christmas I. volcano, etc. (Early Tertiary)	.....
II. Ophiolitic Suite	Foredeep (2)	Geosynclinal	Ophiolitic kindred (Late Cretaceous to Early Tertiary. The age varies in different parts of the region)	Ophiolitic kindred of Geosynclinal Phase (Early Ordovician) Transgression Phase. No igneous activity (Late Ordovician-Silurian)
III. Pacific Suite	Geanticline (3)	Orogenic (*sensu stricto*)	First Orogeny Plutonites: granites etc. (Eocene-Oligocene) Volcanites: basalto-andesites and acid tuffs (Oligo-Miocene) Second Orogeny Plutonites: granites, etc., and basic rocks ("basic front") (Miocene) Volcanites: basalto-andesites and acid tuffs (Mio-Pliocene) Third Orogeny Plutonites: granites, etc. (Quaternary) Volcanites: basalto-andesites and acid tuffs (Plio-Pleistocene)	Early Orogenic Phase Plutonites: foliated granites, oligoclase gneiss etc., migma-rocks; foliated basic and ultrabasic rocks (Late Silurian) Second Orogenic Phase Plutonites: granite, granodiorite, quartz-diorite, gabbro, norite, and peridotite (Late Silurian) Third (?) Orogenic Phase Plutonites: granite, granodiorite, quartz-diorite, etc. with numerous small masses of monzonite, kentallenite, etc. Volcanites: lavas ranging from basaltic andesite to rhyolite (Lower Old Red Sandstone) (Kentallenite etc. and orthoclase-olivine basalts within the above Third (?) Orogenic Phase)
IV. Mediterranean Suite	Backdeep (4)	Late Orogenic	Potash-rich lavas of the Muriah and Ringgit volcanoes (Plio-Pleistocene)	
V. Tholeiitic Plateau Basalts	Hinterland (5)	Post-Orogenic	Fissure eruptions giving rise to over-saturated basalts (tholeiitic basalts) (Quaternary)	Post-orogenic Phase Volcanites: Trachybasaltic kindred (Carboniferous-Early Permian) Quartz dolerite dikes and sills (Earliest Permian)

orogenic or kratogenic regions of the earth's crust.

While these ideas in their present form have influenced and advanced petrogenetic theory, they have never been subjected to the detailed criticism which, in my opinion, would add considerably to their usefulness. Kennedy, for example, practically rules out volcanic rocks from his plutonic association, and one may wonder how the ophiolitic (spilitic) suite and the pyroxene andesite to rhyolite suite fit into his scheme. He does not mention the ophiolitic suite at all and only refers to the basalt-andesite-rhyolite suite in connection with his two postulated lines of basaltic descent (1938, p. 34, 37). Both suites are of worldwide distribution in orogenic zones of all ages, and occur in great bulk (Tyrrell, 1937, p. 91). The ophiolitic suite is recognized by almost all petrologists as being connected with the geosynclinal stage of the diastrophic cycle, and its eruption is the igneous event which corresponds to the opening stage of that series of tectonic events which culminates in the orogenic phase. Scheumann (1924, p. 8) indeed calls it *prototectonic*. The typical basalts, spilites, gabbros, and peridotites (serpentines) are never found along with the rocks of the basaltic lines of descent which are characteristic of nonorogenic regions. In the writer's opinion their provenance, distribution, and association with orogenic events places them indubitably with the plutonic association.

In regard to the basalt—pyroxene andesite—hornblende-mica andesite—dacite—rhyolite series, which appears to be characteristic of orogenic zones and times, basalt is just as subordinate in this series as it is predominant in the volcanic association. Pyroxene andesites are dominant in this suite, and basic pyroxene andesites often rich in hypersthene are the most abundant lavas erupted from the present-day volcanoes of the entire circum-Pacific orogenic zone. There is reason to believe that acid lavas and tuffs are nearly as abundant as the andesites in some regions as, for example, the "cataclysmal eruptions" of acid tuffs, thrice repeated, in the orogenic phases of the Tertiary Indonesian cycle. Fenner (1948, p. 890–892) states that, while the normal lavas of the Arequipa (Peru) region are predominantly andesites, eruptions of rhyolitic tuff reach a thickness of 1000 feet and cover thousands of square miles, and are thus clearly a most important feature of Andean volcanicity. He suggests, in consequence, that the prevalent idea of an overwhelming excess of andesitic lava in the Andean region may require modification. The same relations probably hold in the Katmai region of Alaska and in the North Island of New Zealand. These facts are clearly at variance with the characters Kennedy regards as typical of the volcanic association

While in some passages Kennedy appears to exclude the possibility of his volcanic association occurring in connection with plutonic masses, and cites certain facts as "suggestive of the lack of intimate connection between the plutonic and volcanic associations", he goes on to say (Kennedy and Anderson, 1938, p. 29):

> "This conclusion appears to find support also in the observed relations of batholiths to their roofs, and Bucher [1933] in particular has emphasised the fact that the large acid plutonic masses of the orogenic belts are capable of penetrating the crust to points very near the surface without bursting their roofs in catastrophic fashion. Under such circumstances, moreover, they do not appear to give rise to surface eruptions but actually *cut the effusive rocks of the same igneous episode* [italics present writer's]. This universal absence of lavas belonging to a period contemporaneous with the rise of batholithic intrusions to their highest levels in the crust is of considerable importance".

But, on Kennedy's own showing, it is possible for effusive rocks of the *same igneous episode* to reach the surface *before* the batholiths arrive at their highest levels in the crust. Moreover, as in the Scottish Caledonian and the Indonesian Tertiary orogenic phases, the lavas which are cut by the *final* intrusive efforts of the batholiths are invariably those of the andesite—dacite—rhyolite series,—a series parallel in chemical composition to the rocks of the quartz diorite—tonalite—granodiorite—granite series, *i.e.* the typical rocks of the plutonic association.

Parallelism in chemical composition between the plutonic and volcanic suites is shown by comparison of average analyses of the various types. For example, the average composition of hypersthene andesite which the writer has computed from 114 analyses of rocks called

hypersthene andesite, and pyroxene andesite containing hypersthene, culled from the literature on Recent and Tertiary volcanoes and lava fields of the circum-Pacific region (including the East and West Indies), and certain European fields (Hungary, Sardinia, Aegean Sea) (Tyrrell, 1931b, p. 195), is extraordinarily close to the average of 125 diorites and quartz diorites computed by Daly and to the average of the 55 quartz diorites which are included in the above (Daly, 1933, p. 15–16). Also, comparison of the average analyses of 40 granodiorites and 90 dacites (Daly, 1933, p. 15) and of 546 granites and 126 rhyolites (Daly, 1933, p. 9) strongly reinforces the parallelism.

The lava series ranging from pyroxene andesite to rhyolite, therefore, has the same chemical composition, the same geological and geographical distribution, the same tectonic environment, and appears at the same stage of the tectono-igneous cycle as the plutonic series ranging from quartz diorite to granite. If the latter belongs to the plutonic association there seems to be no escape from the conclusion that the andesite—dacite—rhyolite series likewise does. Indeed if the volcanic association consists predominantly of basaltic lavas with their differentiates, together with subordinate coarse-grained intrusions of "plutonic" type, there is no reason why the predominatingly coarse-grained, more or less "acid", subjacent intrusions of the plutonic association should not give rise to subordinate series of volcanic rocks corresponding to them in chemical range. The writer suggests that the whole trouble with plutonic and volcanic associations is that they are wrongly named. As Kennedy has well shown there are certainly two strongly contrasted series of igneous rocks; but the two groups are contrasted not so much in their plutonic or volcanic characters, or in their chemical compositions, as in their tectonic associations. One group is associated with orogenetic processes during a period of active orogeny in an orogenic zone; the other is associated with epeirogenetic movements in stable nonorogenic regions (kratogens). Kennedy's concept would gain greatly in clarity and usefulness if the plutonic and volcanic associations were renamed orogenic and kratogenic, respectively.

Nature and Origin of Igneous Cycles

At this stage the writer does not attempt to discuss fully the views of the many workers in this field of study, but merely to present in brief his own views, which have been partly adumbrated in earlier pages of this paper, and compare them with a few selected from the abundant literature.

Table 2.—Tectono-Igneous Cycle

Diastrophism	Kindreds	Locus
I. Geosynclinal Phase	1. Ophiolitic Kindred	In Orogen
II. Orogenic Phase (with two or three subphases)	2. Granodiorite-Andesite Kindred	In Orogen
III. Post-Orogenic Phase (with two subphases)	3. Trachybasaltic Kindred 4. Quartz dolerite Kindred	In Kratogen

For the highly complicated region of Northwestern Europe, consisting, as it does, of three ancient orogens welded onto the Scandinavian-Baltic Shield, the complete tectono-igneous cycle worked out for the Scottish Paleozoic, which may be regarded as typical for the above region, is shown in Table 2, reduced to its simplest terms and using the simple diastrophic cycle already given.

The cycle begins with an ophiolitic kindred associated with the geosynclinal or prototectonic stage of orogeny, and confined to the orogen. At the climax of orogeny, after a long period of comparative quiescence in diastrophic and igneous activity, vast irruptions of the granodiorite-andesite kindred take place, and may reach two or three peaks of intensity preceding, during, and immediately subsequent to the folding. In the earliest subphase processes of palingenesis, anatexis, and granitization transform the roots of the geosynclinal prism into plutonic rocks (Read's sense). In the later subphases the completion of mobilization makes available magmas capable of injection

into the upper brittle parts of the crust, where they solidify to form igneous rocks of the granodiorite-andesite kindred. Finally, toward the end of the cycle, when the orogen has been united with the adjacent kratogen and then participates in the ensuing kratogenic tectonic reactions (block faulting, etc.), there are abundant local eruptions of the trachybasaltic kindred, followed by an extensive series of dikes and sills of the quartz dolerite kindred. The first and second tectono-igneous phases are products of orogenic, the third and fourth of kratogenic activity.

This cycle of tectono-igneous events is probably too neat to be the whole truth, but Table 3 shows that it holds fairly generally throughout the eras within the western half of Europe, and will possibly be found to hold for eastern North America. The third column of Table 3 covers igneous events of the Alpine cycle throughout the Mediterranean and Alpine region as far as the Balkans and Turkey. Table 1 shows that the above theoretical tectono-igneous cycle applies to the Indonesian Tertiary rather closely; and we may surmise, therefore, that it fits tectono-igneous events throughout the whole length of the Alpine-Himalayan system. In the third column of Table 3 the writer has omitted the enormous North Atlantic region of flood basalts and the accompanying local centers. This is because he believes it to be a thalattogenic event

TABLE 3.—COMPARISON OF THE IGNEOUS PHASES OF THE CALEDONIAN, HERCYNIAN, AND ALPINE OROGENIC CYCLES

Diastrophic Cycle* → Caledonian (Scotland)		Hercynian (Western Europe, Etc.)	Alpine (Central Europe and Mediterranean)*
III. Post-orogenic Phase	4. Quartz dolerite kindred; dikes and sills (Earliest Permian)	4. Newark basalts, Palisade sill, etc. of eastern North America (Triassic)	.....
	3. Trachybasaltic kindred; lavas and associated intrusions (Lower Carboniferous to Early Permian)	3. ? Exeter lavas, some potash-rich types (Middle to Upper Permian)	3. Trachybasaltic kindred of Extra-Alpine regions (Late Tertiary to Recent)
II. Orogenic Phase	2b. Granodiorite-andesite kindred; plutons and lavas (Late Silurian to Lower Devonian)	2b. "Porphyrite" and "Melaphyre" lavas of western Germany (Permian) Granite and granodiorite plutons of Southwest England and Germany (Late Carboniferous to Permian)	2b. Andesite fields, Hungary and Balkans (Tertiary) Granodiorites, tonalites, etc. Adamello & Tonale Alps, etc. (Late Cretaceous)
	2a. Granodiorites and granites; Gneisses, migmarocks, injection complexes (Late Silurian?)	2a. Granodiorite and granite syntectonic intrusions, gneisses, and granulites (Late Paleozoic)	2a. Synorogenic granodiorite gneisses, etc. (Middle Cretaceous to Paleocene—long believed to be Hercynian)
I. Geosynclinal Phase	1. Ophiolitic kindred (Lower Ordovician)	1. Ophiolitic kindred (Middle Devonian to Lower Culm)	1. Ophiolitic kindred (Mesozoic—various ages)

* Most of the data for this column derived from Burri and Niggli (1945)

connected with the formation of the North Atlantic Ocean, entirely alien to, although approximately contemporaneous with, the tectono-igneous events of the Alpine diastrophic cycle. So also with the Deccan Traps which are considered to be similarly connected with the inbreaks which formed the northern part of the Indian Ocean.

The ultimate origin of the kindreds that the writer has associated with the tectono-igneous cycle, and many others, is still largely a matter of conjecture and speculation. Umbgrove (1939) has provided a valuable review of some of the ideas that have been put forward. Many authors have pointed out that basic and ultrabasic rocks develop in geosynclines at the beginning of the tectono-igneous cycle, and this is perhaps the best-established generalization in the subject (Holmes, 1932; Bucher, 1933; Rittmann, 1936). The writer subscribes to Hess's views (1938) on the origin of geosynclinal ultrabasic rocks, and would extend them to explain the ophiolitic suite of which serpentine in particular is such an important constituent. In the development of a geosyncline the lower parts of the crystalline crust are buckled downward with the result that the basaltic layer and even the ultrabasic material at greater depths enter regions of higher temperature with the formation of sheet-like magmatic pools at various levels which may ultimately reach shallower positions in the crust along planes of tectonic weakness (Kennedy and Anderson, 1938, Fig. 1, p. 39; Hess, 1938, Fig. 1, p. 334). Sir E. B. Bailey and W. J. McCallien have even announced the discovery of serpentine *lavas* in Anatolia (1950). This may explain the irruptions of basic and ultrabasic rocks along the geosynclines; but there is still no adequate explanation of the frequent excess of soda resulting in the characteristic spilites and keratophyres of the ophiolitic phase which remains an unsolved petrological mystery.

I cannot entertain Van Bemmelen's tentative hypothesis (1950, p. 218) that "the ophiolitic suite is a regional low-level basic front" representing a "geochemical culmination of cafemic constituents in front of acidification and migmatization of the crustal base", for the sufficient reason that, in all the tectono-igneous cycles yet investigated, the ophiolitic phase is separated by a long period of quiescence from the granitization processes which set in during the opening stages of active folding in the orogen. In many regions for which granitization has been advocated it is these very rocks of the ophiolitic kindred, and especially the graywackes with which they are invariably associated, which have been affected by that process.

Another point which seems to be of considerable importance in this connection is the unexpected fact that the order of size of ophiolitic accumulations is little smaller than that of the flood basalts. Extrapolating from the volume of the ophiolitic suite of the Caledonian geosyncline in southern Scotland (Tyrrell, 1937, p. 91), which is estimated at 1000 cubic miles for every 100 miles along the strike, it is possible that 30,000 cubic miles of basaltic and spilitic lavas were erupted along the Caledonian orogen in Northwestern Europe; and as there was almost certainly an equal quantity of basic and ultrabasic intrusive rocks associated with the lavas the total amount of irrupted material may have been of the order of 60,000 cubic miles. If this estimate is acceptable the enormous bulk of the ophiolitic suite in most old fold-mountain ranges may be regarded as more favorable to the idea of its rapid derivation from elongated reservoirs of fused material of basic and ultrabasic composition along the geosynclinal zones rather than as a "basic front".

In regard to the Orogenic Phase the writer has shown that what he has called the granodiorite-andesite kindred is the characteristic igneous product. The writer believes with Eskola (1932) that, in orogens, processes of palingenesis and anatexis with assimilation and differential melting, to which we may now legitimately add the Transformist mechanisms of granitization, are responsible for the ultimate appearance of magmas of the granodiorite-andesite kindred in the upper levels of the crust. Von Bubnoff's fine diagram of the Schwarzwald and Odenwald tectono-igneous levels, reproduced with slight modification in the writer's *Principles of petrology* (1926a, p. 335), perfectly illustrates the point of view. In kratogens, on the other hand, the writer believes that granite (and usually a more alkalic variety of granite than that of the oro-

gens) is due to processes of differentiation (Tyrrell, *in* Eskola, 1932, p. 481); but, of course, processes of differentiation may also take place in the later stages of the orogenic phase when true magmas have been generated and are held for a time in subcrustal reservoirs. All the above-mentioned processes co-operate in generating "granitoids"[9]. This is, of course, eclecticism; but the writer would rather err in that direction than the opposite one, in which the inventors of a new and promising theory, in the first flush of enthusiasm, overextend its legitimate field of application to the exclusion of all others in the attempt to make one theory alone responsible for the whole range of tectono-igneous phenomena.

There are at least two, and often more, peaks of intensity during the orogenic phase of tectono-igneous activity. In the Caledonian of Scotland and the Tertiary cycle in Indonesia there seem to be three; and in the Argentinian Andes Backlund (1926) distinguishes no fewer than five. The first sub-phase, which is pre- or syntectonic, is marked by the formation of gneisses, migma rocks, and injection complexes mainly of "acid" composition, although more basic types may participate to varying extents. This is the period in which palingenetic and anatectic processes, innfelting and differential melting, assimilation, and granitization in its narrower sense reach their highest intensity in the deep-seated theaters of operation at the bases of the geosynclinal prisms with which marginal parts of the adjacent forelands are also involved. As time goes on true magma is generated from the mixed materials as the mobilization admitted by all but the most extreme Transformists proceeds. Magma derived from these processes reaches higher levels in the crust during later and less intense spasms of tectonic activity and gives rise to granitoid plutons with approximately contemporaneous outbursts of andesitic and rhyolitic volcanoes. How these bodies are emplaced or erupted is admittedly difficult to explain, but as these vexed questions, however important, are largely irrelevant to the topic of distribution in space and time, the writer does not propose to enter upon them here.

[9] *Granitoid*—an omnibus term for the series of granite-granodiorite (adamellite, etc.)-tonalite-quartz diorite, etc.

Apparently it is only the later and less powerful paroxysms of tectonic activity which are associated with volcanic eruptions of the granodiorite-andesite kindred accompanied or followed by the emplacement of large granitoid plutons. This fact consorts well with R. T. Chamberlin's distinction between thick-shelled and thin-shelled fold mountains (1919). In the thin-shelled type such as the Pennsylvania Appalachians, the Canadian Rockies, and the Alps, a relatively thin superficial portion of the crust is characterized by intense folding, thrusting, and dynamic metamorphism, with a maximal amount of shortening and little or no volcanism. On the other hand thick-shelled mountains such as the Colorado Rockies and much of the Andes are marked by comparatively gentle open folding, moderate shortening affecting a relatively thick zone of the crust, and strong epeirogenic uplift with normal faulting which may likewise reach and tap considerable depths. It is with the latter type that the granodiorite-andesite kindred, in both its volcanic and plutonic manifestations, is connected. The same contrast may be found in one and the same orogenic zone as, for example, in the Colorado Rockies (thick-shelled) and the Rockies of Montana and Alberta (thin-shelled). The intense compression suffered by thin-shelled mountains appears to be inhibitory of surface volcanism.

When the orogenic belt has been welded onto the adjacent kratogen the very different conditions of the Post-Orogenic (Continental) Phase supervene. The diastrophism is of epeirogenic character; great faults and zones of weakness penetrate deeply into the continental basement, and may tap pools of magma which have been formed within both the intermediate layer (layer of olivine-free or tholeiitic basalt, Kennedy and Anderson, 1938, p. 38) and the olivine basalt layer of the earth. Kennedy and Anderson (1938) have investigated the thermal and tectonic conditions under which such pools may be formed, and Figure 1 of their paper provides a succinct statement of their views. On the other hand it is arguable that the olivine-free or tholeiitic basalt layer does not exist as such, but that magma pools of that composition are formed as the result of interaction between activated olivine basalt and the more acid underside of the continental

crust. Whichever view is adopted it is certain that both types of basaltic magma become available in great quantity at the Post-Orogenic (Continental) Phase of the tectono-igneous cycle.

According to the conception of the tectono-igneous cycle that is developed in this paper the Continental Phase is characterized by the appearance of a trachybasaltic kindred and a quartz doleritic or tholeiitic kindred. Whereas in the Caledonian tectono-igneous cycle the trachybasaltic kindred largely precedes the quartz dolerite sills and dikes, this order in time does not necessarily hold for other cycles, nor does it hold for the basalt floods and associated intrusions connected with the major inbreaks by which parts of the oceans come into existence. There is, in fact, plenty of evidence for the simultaneous availability of the two kindreds. Both the North Atlantic basalt region and the Deccan traps of India contain rocks assignable to the olivine basalt line of descent (trachybasaltic kindred) and the tholeiitic basalt line of descent (quartz dolerite or tholeiitic kindred). In the North Atlantic region the general order of magmatic production is trachybasaltic followed by tholeiitic; but, while there are many exceptions to this rule, there are also later magmatic resurgences which do follow the above general order (Tyrrell, 1949, p. 419). The petrography of the Deccan traps is not yet fully known, but, from the literature, they seem to belong mainly to the tholeiitic affiliation. Nevertheless, dikes and ring intrusions of alkalic types cutting and doming the Deccan traps are known in the Kathiawar Peninsula (Mathur *et al.*, 1926). This seems, therefore, to represent a reversal of the more usual order.

The tectono-igneous cycle which the writer has endeavored to expound is based on the complicated, highly disturbed, and perhaps rather abnormal western part of the Eurasian continent. In conclusion the writer expresses the hope that American petrologists will now take up the task of setting forth the tectono-igneous cycle in North America which, in many respects, is the type continent of the world.

References Cited

Anderson, J. G. C., 1947, The geology of the Highland Border: Stonehaven to Arran: Royal Soc. Edinburgh Trans., v. 61, pt. 2, p. 479–515.

Backlund, H. G., 1926, Magmatic activity and mountain folding in the Andes of South Mendoza: Geol. Mag., v. 63, p. 410–422.

Bailey, E. B., 1928, The Paleozoic mountain systems of Europe and America: British Assoc. Adv. Sci. Sect. C, Glasgow, p. 57–76.

Bailey, E. B., and Maufe, H. B., 1916, The geology of Ben Nevis and Glen Coe: Geol. Survey Scotland Mem., 247 p.

Bailey, E. B., and McCallien, W. J., 1950, The Ankara Mélange and the Anatolian thrust: Nature, v. 166, p. 938–940.

Bemmelen, R. W. van, 1949, The geology of Indonesia, vol. 1A: The Hague, Govt. Printing Office, 732 p.

—— 1950, On the origin of igneous rocks in Indonesia: Geol. en Mijnb. 12, no. 7, p. 207–220.

Billings, M. P., and Keevil, N. B., 1946, Petrography and radioactivity of four Paleozoic magma series in New Hampshire: Geol. Soc. America Bull., v. 57, p. 797–828.

Bubnoff, S. von, 1923, Der Gliederung der Erdrinde: Fortschr. Geol. Palaeont., 3, 84 p.

—— 1924, Die Gebirgsbau Europas: Geol. Rundschau, Bd. 15, p. 147–174.

Bucher, W. H., 1933, The deformation of the earth's crust; Princeton Univ. Press, 518 p.

Burri, C., and Niggli, P., 1945, Die jungen Eruptivgesteine des mediterranen Orogens. I. Hauptteil. Einfuhrung-Die Ophiolithe-Allgemeines über das Verhalten basischer Magmen: Vulkaninst. Immanuel Friedlander, Zurich, Pub. no. 3, 654 pages.

Chamberlin, R. T., 1919, Building of the Colorado Rockies: Jour. Geology, v. 27, p. 148–164, 225–251.

Cross, Whitman, 1912, Petrological abstracts and reviews: Jour. Geology, v. 20, p. 326–372.

Daly, R. A., 1933, Igneous rocks and the depths of the earth: N. Y., McGraw-Hill Book Co., 2d ed., 598 p.

Dewey, H., and Flett, J. S., 1911, On some British pillow-lavas and the rocks associated with them: Geol. Mag., v. 48, p. 202–209, 241–248.

Eskola, P., 1932, On the origin of granitic magmas: Min. u. Petr. Mitt, Bd. 42, p. 455–481.

Fenner, C. N., 1948, Incandescent tuff flows in southern Peru: Geol. Soc. America Bull., v. 59, p. 879–893.

Goldschmidt, V. M., 1916, Ubersicht der Eruptivgesteine im Kaledonischen Gebirge zwischen Stavanger und Trondhjem: Vidensk. Skrift. I. Math.-Nat. Kl., Kristiania, no. 2, 140 p.

Grout, F. F., 1948, *in* Gilluly, James (*Chairman*), Origin of granite: Geol. Soc. America Mem. 28, p. 45–54.

Harker, A., 1911, Some aspects of modern petrology: British Assoc. Adv. Sci., Pres. Address, 12 p.

Henderson, S. M. K., 1935, Ordovician submarine disturbances in the Girvan district: Royal Soc. Edinburgh Trans., v. 58, p. 487–509.

Hess, H. H., 1938, A primary peridotite magma: Am. Jour. Sci., v. 35, p. 321–344.

Holmes, A., 1928, Radioactivity and earth movements: Geol. Soc. Glasgow, v. 18, p. 559–606.
—— 1932, Origin of igneous rocks: Geol. Mag., v. 69, p. 543–558.
Judd, J. W., 1886, On the gabbros, dolerites and basalts of Tertiary age in Scotland: Geol. Soc. London Quart. Jour., v. 42, p. 49–89.
Kennedy, W. Q., and Anderson, E. M., 1938, Crustal layers and the origin of magmas: Bull. Volc. Naples, Ser. 2, T. 3, p. 23–82.
Kober, L., 1921, Der Bau der Erde: Berlin, Gebrüder Borntraeger, 2d ed. 1928, 324 p.
Krumbein, W. C., and Sloss, L. L., 1951, Stratigraphy and sedimentation: San Francisco, W. H. Freeman and Co.
Krynine, P. D., 1941, Paleogeographic and tectonic significance of graywackes (Abstract): Geol. Soc. America Bull., v. 52, p. 1916.
MacGregor, A. G., 1948, Problems of Carboniferous-Permian volcanicity in Scotland: Geol. Soc., London Quart. Jour., v. 104, p. 133–153.
Mathur, K. K., Dubey, V. S., and Sharma, N. L., 1926, Magmatic differentiation in the Girnar Hills: Jour. Geology, v. 34, p. 289–307.
Niggli, P., 1921, Lehrbuch der Mineralogie: Berlin, Borntraeger, 694 p.
Pettijohn, F. J., 1943, Archaean sedimentation: Geol. Soc. America Bull., v. 54, p. 925–972.
Ramsay, W., 1924, The probable solution of the climate problem in geology: Geol. Mag., v. 61, p. 152–163.
Read, H. H., 1931, The geology of central Sutherland: Geol. Survey Scotland Mem., 238 p.
—— 1948, *in* Gilluly, James (*Chairman*) Granites and granites: Geol. Soc. America Mem. 28, p. 1–19.
Read, H. H., and MacGregor, A. G., 1948, The Grampian Hills: Great Britain Geol. Survey and Mus., 2d ed., Edinburgh, 83 p.
Reynolds, D. L., 1942, The albite-schists of Antrim and their petrogenetic relationship to Caledonian orogenesis: Royal Irish Acad. Proc., v. 48, p. 43–66.
Richey, J. E., 1939, The dykes of Scotland: Edinburgh Geol. Soc. Trans., v. 13, pt. 4, p. 393–435.
Rittmann, A., 1936, Vulkane und ihre Tätigkeit: Stuttgart, Ferdinand Enke, 186 p.
Scheumann, K. H., 1924, Prävariskische Glieder der Sächsisch-Fichtelgebirgischen Kristallinen Schiefer I. Die magmatisch-orogenetische Stellung der Frankenberger Gneisgesteine: Abh. d. Math.-Nat. Kl. Sächs. Akad. Wiss. Bd. 39, 61 p.
Sonder, R. A., 1922, Die erdgeschichlichen Diastrophismen im Lichte der Kontraktionslehre: Geol. Rundschau, Bd. 13, p. 217–272.
Steinmann, G., 1905, Die geologische Bedeutung der Tiefseebildungen unde der ophiolitischen Eruptiva. Geologische Beobachtungen in den Alpen II: Ber. naturf. Ges. Freiburg, Bd. 16, p. 44–65.
—— 1926, Die ophiolitischen Zonen in den mediterranen Kettengebirgen: 14th Internat. Geol. Cong., Madrid, fasc. 2, p. 637–667.
Tyrrell, G. W., 1914, A petrographical sketch of the Carrick Hills, Ayrshire: Geol. Soc. Glasgow Trans., v. 15, pt. 1, p. 64–83.
—— 1926a, The principles of petrology: London, Methuen, 349 p.
—— 1926b, Igneous action and earth movement in Scotland and Scandinavia: Geol. Mag., v. 63, p. 284–286.
—— 1931a, Volcanoes: London, Home Univ. Lib., 256 p.
—— 1931b, The South Sandwich Islands; Report on rock specimens: Discovery Repts., v. 3, p. 191–198.
—— 1933, Greenstones and greywackes: Compte Rendu Réunion Internat. pour l'étude du Precambrian et des vieilles châines de Montagnes, Finland, 1931, Helsinki, May 1933, p. 24–26.
—— 1937, Flood basalts and fissure eruption: Bull. Volcanologica, Naples, ser. 2, tome 1, p. 88–111.
—— 1939, Relations of igneous activity to tectonics: 17th Internat. Geol. Cong., Moscow, 1937, vol. 1, p. 213–217.
—— 1949, The Tertiary igneous geology of Scotland in relation to Iceland and Greenland: Medd. Dansk. Geol. For. 11, p. 413–440.
Umbgrove, J. H. F., 1939, The relation between magmatic cycles and orogenic epochs: Geol. Mag., vol. 76, p. 444–450.
Van Hise, C. R., and Leith, C. K., 1911, The geology of the Lake Superior region: U. S. Geol. Survey Mon. 52, 641 p.
Walker, F., 1935, The late Palaeozoic quartz-dolerites and tholeiites of Scotland: Min. Mag., vol. 24, p. 131–159.
Williams, C. R., and Billings, M. P., 1938, Petrology and structure of the Franconia quadrangle: Geol. Soc. America Bull., v. 49, p. 1011–1044.
Wolff, F. von, 1914, Der Vulkanismus: Stuttgart, F. Enke, Bd. 1, 711 p.

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PROVENIENCE OF PYROCLASTIC MATERIALS

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ABSTRACT

Recent studies of rhyolitic and pyroclastic materials, and in particular of welded tuffs and bentonites, show that they occur over wide areas and in volumes which greatly exceed earlier evaluations. Volcanic ash and bentonite occur in the eastern United States where such materials were long unrecognized. In most of the western States, only preliminary studies have been made, but they indicate the presence of these materials in volumes measured in thousands of cubic miles. Less is known about pyroclastic materials in most other parts of the world, but local studies in New Zealand, Australia, and the Dutch East Indies show immense volumes of pyroclastic materials.

CONTENTS

TEXT

	Page
Introduction	427
Ash falls	428
Bentonite beds	428
Tuffs	430
Dispersal of volcanic dust	432
References cited	433

ILLUSTRATIONS

Figure	Page
1. Map showing distribution of pyroclastic materials east of the Rocky Mountains	430

INTRODUCTION

This paper describes the mode of occurrence of the important types of rhyolitic and pyroclastic materials and illustrates by means of a number of outstanding deposits their age, extent, and volume. Not all the recognized occurrences can be discussed in this paper, and, even if this were possible, it would not give an adequate presentation of these materials in the western United States. Numerous areas of them are known but have received only superficial examination.

The petrology and genetic relations of volcanic rocks, including their pyroclastic equivalents, have long interested geologists. Nevertheless, petrographic relations have been stressed more than the extent of the pyroclastic rocks. Thus the great volume of these materials has not been fully recognized. There are several reasons for this; the most important is that in the past pryoclastic materials have commonly not been recognized as such.

Of all the types of igneous rocks, those of pyroclastic origin are the least amenable to quantitative evaluation. The intrusive rocks, and even the extrusive ones, leave at least a partial record. In contrast, ash and many tuff deposits are subject to various vicissitudes. Ease of erosion, wide dispersion, thinning out to inconspicuous lenses, admixture with detrital materials, and permeability leading to ease of alteration all tend to obscure or destroy the record of their existence. However, several lines of study are adding to our knowledge of pyroclastic materials and proving their wide occurrence and great importance.

This paper considers the provenience of the pyroclastic materials allied to rhyolites, and so it must be understood what is included in the grouping. For the present purpose dacites and quartz latites, and probably part of the latites, may be grouped with rhyolites since the geologic relations of the pyroclastic phases of these rock types are similar. However, rhyolites would greatly exceed all other rock types in such a grouping, since quartz latites more commonly form flows than do rhyolites. This excludes the andesitic ash materials, although they have many characteristics in

common with rhyolitic ash eruptions. However, andesitic ash materials constitute only a small part of andesitic rocks as a whole, whereas rhyolitic ash makes up a very large part of rhyolitic rocks.

Ash Falls

Ash that falls on land or directly into the ocean commonly leaves a very incomplete record. Where ash falls on land it is subject to rapid erosion and may be removed from the geologic record. Thick deposits of coarser tuffs are somewhat more persistent. If ash deposits are thick or extensive they may be redeposited in local lakes or marine embayments and leave evidence of their existence, but the original amount would normally be problematical. The transitory character of much ash-fall material is illustrated by the Krakatoa eruption of 1883; Verbeek (1886, p. 140) estimated a volume of about 4.3 cubic miles. The volume of ash fall material at Katmai has been estimated by Griggs (1922, p. 29) to be between 4 and 5 cubic miles, but that estimate does not include the "sand flow" which was estimated (Griggs, 1922, p. 253) to have had a volume of about a cubic mile.

Swineford and Frye (1946) have described volcanic ash from several horizons associated with beds of Pliocene and Pleistocene age in Texas, Oklahoma, Kansas, and Nebraska. The area in which these occur extends northeast as far as Iowa, from the probable source in New Mexico, and covers about 300,000 square miles. Local deposits range up to 6½ feet in thickness, but the average thickness, though unknown, must be much less, and the volume, although unknown, must be large. These occurrences illustrate the way ash may (1) fall into the sea and be largely lost to the record; (2) fall upon land only to be promptly eroded away; or (3) be left in such scattered deposits that volumes are indeterminable. Throughout geologic time erosion must have largely destroyed the record of similar eruptions.

Bentonite Beds

Within recent years the stratigraphic and economic interest in bentonite has led to a large fund of knowledge about this clay material, derived from volcanic ash. Bentonite commonly occurs as clay beds in a sedimentary sequence, may differ little from other clays in superficial characteristics, and its beds commonly form no conspicuous outcrops. Hence beds of bentonite are commonly recognizable only after intensive study, and new occurrences are continually being found. Buried beds of bentonite have been recognized by means of drill holes, and so are best known in oil fields. Drill holes commonly fail to reveal their total lateral extent, and so the approximately determined volume may be greatly exceeded by the actual volume.

A very widespread occurrence of a clay material derived from volcanic ash is represented by several beds in Ordovician strata in the eastern United States, and extending into Canada. This material was first recognized at Highbridge, Kentucky, by Nelson. It has been recognized in Kentucky, Tennessee, Alabama, and Virginia by Nelson (1922); in Pennsylvania by Bonine and Honess (1929); in the Georgian Bay region of Canada by Maddox (1930); near St. Louis, Missouri, by Allen (1929); and in northeastern Iowa and near-by parts of Wisconsin and Minnesota by Allen (1929; 1932). Kay (1931) has studied the material in New York and many of the aforementioned localities. Thus the area included has diameters of 600 to 900 miles and covers 300,000 to 400,000 square miles. Nelson estimated the volume to be 66 cubic miles, but since then the known area has been greatly extended, and the volume must be several times that figure.

The ash materials of the Ordovician consistently become coarser to the southwest, and one horizon at the Chickamagua dam in Tennessee is composed dominantly of feldspar grains of variable size and biotite plates, ranging up to more than a millimeter in diameter.

Bentonite has been recognized in Ordovician strata in Poland, north Germany, and southern Sweden.

A thin clay bed derived from volcanic ash has been recognized by Hass (1948) in the Upper Devonian beds of Tennessee, and it is believed to extend far to the east.

The Permian basin of Texas and New Mexico is characterized by at least five bentonite beds whose maximum aggregate thickness is 10 feet. Since the relations are known almost wholly

from drill holes, the full extent is not definitely known. However, W. B. Lang (personal communication) points out that the average thickness is at least 2 feet, and the area over which they occur is probably not less than 70,000 square miles. This indicates that the recognizable volume of bentonite in this basin is of the order of 27 cubic miles.

No attempt is made to list or estimate the volume of bentonite beds that occur throughout the mountain region of the western United States. Many individual localities are known, but these would give but a spotty picture of their extent, and not even an approximate idea of their great volume.

The Gulf Coastal Plain is now known to be a region of major volcanic activity. Pyroclastic materials have been recognized in Alabama, Mississippi, Louisiana, Arkansas, Oklahoma, and Texas, and they probably occur also in Florida. They are also found bordering the Coastal Plain area of Mexico; and a thin deposit has been recognized by Stevenson (1936, p. 489) in New Jersey. Volcanic materials have not been recognized in the intervening regions of North Carolina, South Carolina, Virginia, and Maryland, but a connection and a former continuity with the Gulf Coast areas seems probable. Eruptive activity was long continued; volcanic materials are abundant in the Upper Cretaceous strata and occur throughout the Eocene deposits and in deposits as young as Miocene.

Volcanic materials are abundant in the region near Nashville, Arkansas, where several eruptive centers are believed to be located (Ross, Miser, and Stephenson, 1930, p. 189). Volcanic materials, admixed with sedimentary rock fragments, are abundant in the Upper Cretaceous rocks for 100 miles to the west in eastern Oklahoma and northeastern Texas. Similar materials have been recognized in wells far to the south in Texas and Louisiana, in one instance at a depth of 11,418 feet. Moody (1949, p. 1418) states

> "Volcanism contemporaneous with the accumulation of some of the Upper Cretaceous formations of the northern Coastal Plain is made evident by the recognition, in samples from wells dispersed throughout the entire province, of fragments of volcanic rocks . . . .
>
> "In addition to these occurrences, which characterize virtually every set of well samples obtained in a large area embracing positions of east Texas, south Arkansas, north Louisiana, and central Mississippi, it is found that certain wells have penetrated noteworthy thicknesses of tuffs and possibly of volcanic agglomerates."

In the Monroe, Louisiana, gas field 1500 feet of tuff was encountered in wells. In Humphreys County, Mississippi, 2035 feet was penetrated without reaching normal sediments. Similar thicknesses were encountered in wells at Jackson, Mississippi.

L. W. Stephenson (personal communication) has recognized bentonite and other pyroclastic materials at numerous localities in Texas in the Upper Cretaceous Eagle Ford formation, Taylor marl, and Navarro group. West of San Antonio a bed of bentonite grading up into admixed detrital materials reaches a thickness of 50 feet.

Volcanic ash beds have been described from southwestern Texas by Bailey (1926). These extend for about 240 miles from Moulton, Lavac County, almost to the Rio Grande near Rio Grande City. The total volume is estimated by Bailey (1926, p. 178) to be 150 cubic miles.

In Mississippi numerous bentonite beds are known in the Upper Cretaceous and lower Tertiary strata. Beds aggregating 50 feet are reported by Watson Monroe (personal communication). At Montgomery, Alabama, an important bentonite bed occurs in the Eutaw formation (Upper Cretaceous); it also occurs in the same strata in Georgia. Farther south in Alabama bentonite beds of Tertiary age are known but seem to be thinner and more impure than in Mississippi.

Most of the volcanic centers in the Gulf Coast region from which these pyroclastic materials were derived are deeply buried; the gulfward extent is unknown, the amount removed by erosion is problematical, and a large part of our information (significant but very incomplete) has been gained from oil-well drilling. Nevertheless this shows that the Gulf Coastal Plain is an outstanding volcanic region and contains a vast amount of pyroclastic materials.

The northern Great Plains region is characterized by thick and extensive deposits of bentonite in the Cretaceous formations. M. M. Knechtel (personal communication) states

that these reach a total thickness of 175 feet in north-central Wyoming. They extend eastward to central Nebraska, south into Colorado, and north into central Manitoba, where a single bed is 6–10 inches thick. The areas in which these beds occur cover 300,000 to 400,000 square miles and indicate a volume of 3500 to 5000 cubic miles. How much greater this may have been in the past is unknown. Bentonite beds also occur in the Tertiary rocks of the same region, and those beds may have a volume of the same order of magnitude.

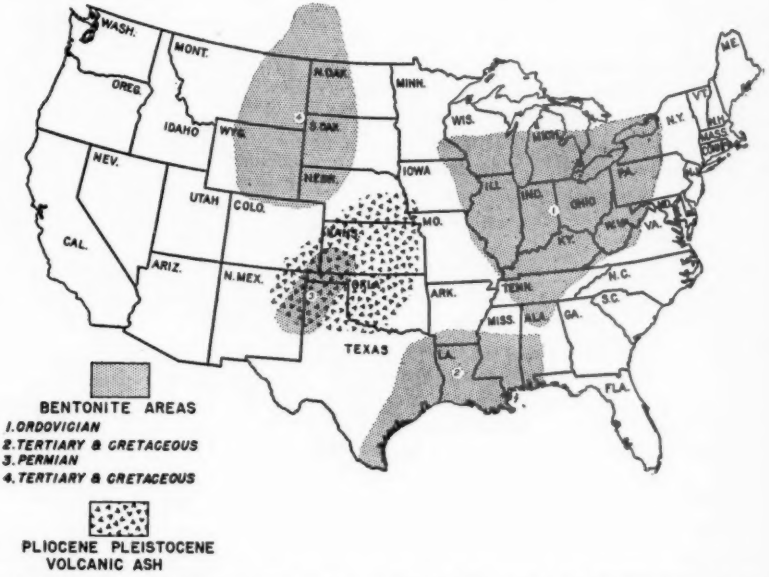

FIGURE 1.—MAP SHOWING DISTRIBUTION OF PYROCLASTIC MATERIALS EAST OF THE ROCKY MOUNTAINS

Figure 1 illustrates the approximate distribution of pyroclastic materials in the United States east of the Rocky Mountains and indicates the wide distribution of these materials in a region where until recently they were believed to be almost absent. They also occur throughout the western United States, but nothing but a large-scale, and very detailed map would represent their distribution.

TUFFS

Tuffs, especially welded tuffs, are the one type of pyroclastic rocks that is most persistent and provide a fair opportunity for evaluation. These rocks are now arousing keen interest, and our knowledge of them is growing rapidly. Many, probably most of the rocks of the western United States previously mapped as rhyolite flows, are now known to be welded tuffs, and such beds are so extensive that a great amount of detailed mapping will be required to reveal their extent. No doubt the same is true for other parts of the world.

The San Juan region of southwestern Colorado, studied by Cross and Larsen (1935), is one of the great volcanic areas of the world. They estimate that the San Juan tuff of that region has a volume of 700 cubic miles. The Silverton volcanic series of the same region has been estimated (Cross and Larsen, 1935, p. 57; W. S. Burbank, personal communication) to have a volume of 250 cubic miles, an important part of which is tuff. Both these tuffs are believed to be of Miocene age. The volcanic rocks of the San Juan region of Tertiary age cover an area of about 12,000 square miles, and a large proportion of the rhyolitic and related rock types are pyroclastic materials.

The writer and Robert L. Smith have been

studying the volcanic rocks, including welded tuffs and related pyroclastic rocks of Pleistocene age in the Jemez Mountain region of New Mexico. The welded tuffs of this region cover more than 350 square miles and have a volume of about 50 cubic miles. Pyroclastic rocks occur in a number of places in the Rio Grande valley from Santa Fe north to the Colorado State line, and the Santa Fe formation (Tertiary) of that region contains numerous beds of volcanic ash. In addition to the detailed studies in the Jemez Mountain region of New Mexico, Ross and Smith have made reconnaissance studies of areas in New Mexico, Colorado, Utah, Nevada, and Idaho. Throughout this entire region rhyolitic welded tuffs were readily recognizable and are present over large areas and in large volume.

Tuff beds, including welded tuffs, have been recognized in West Texas. Tertiary tuffs cover large areas and attain great thicknesses in the Santa Rita and Globe-Superior region of Arizona. J. H. Mackin (report in preparation) has mapped large areas of welded tuffs of Tertiary age in the vicinity of Iron Springs, Utah. These rocks characterize the ridges of this portion of the Basin and Range country and extend west and southwest an unknown distance into Nevada. Callaghan (1939) has identified large areas of welded tuffs of Eocene or Oligocene age in the vicinity of Marysville, Utah.

An examination of the thin sections of rocks studied by Zirkel (1876) as a part of the "Geologic Exploration of the Fortieth Parallel" shows that a large portion of this group of rocks of Nevada and Utah were welded tuffs. The basins between the ridges must contain still more pyroclastic materials.

Welded tuffs constitute much of the rocks of the Yellowstone National Park and seem to be continuous with those of southeastern Idaho. G. R. Mansfield (Mansfield and Ross, 1935, p. 310) reports that the welded tuffs of Pliocene? age of the Idaho area "disappear beneath the Snake River plain and similar rocks emerge again on the north," about 90 miles away. They flank the west slope of the Teton Range and extend to the Red Lake region of southern Montana. The tuffs in the Yellowstone National Park and those of southeastern Idaho seem to constitute parts of the same geologic episode. Kirkham (1931, p. 582) reports great thicknesses of rhyolitic tuffs in the Snake River region, which shows that their probable total volume is many hundreds of cubic miles.

T. P. Thayer (personal communication) has studied large areas of welded tuffs of Pliocene age in central Oregon, and many of the rocks intervening between southeastern Idaho and central Oregon probably represent pyroclastic materials. This suggests that the region extending from northwestern Wyoming to Oregon will eventually prove to contain welded tuffs in the order of thousands of cubic miles.

Gilbert (1938) has described a welded tuff of Pleistocene age in California and estimated the volume to be approximately 35 cubic miles. Bramlette (1946) reports great thicknesses of welded tuffs, bentonite, and volcanic ash in the Monterey formation (Miocene) in California. The thicknesses vary greatly, but, since the Monterey extends from a point north of San Francisco to beyond Los Angeles, the total volume of pyroclastic material, though not definitely known, must be very great.

Important thicknesses of tuff and ash materials occur in the Sierra Nevada of California. In the Coast Ranges and the basins of that State, the areas of these materials are so large and the thicknesses so great that the total volumes are probably of the order of thousands of cubic miles.

Williams (1942) has estimated that pyroclastic materials of Pleistocene age at Crater Lake, Oregon, have a volume of 10 to 12 cubic miles.

The aforementioned welded tuff localities in the western United States represent only localities where notable occurrences have been studied or have been observed but which remain to be studied in detail. Other large areas in the same region have similar geologic relationships and certainly contain other large extents of welded tuffs.

Welded tuffs have been described in a few outstanding localities in other countries, and mentioned but not adequately studied in a number of others. The literature shows that, in many localities of the world, geologists included descriptions of rhyolitic rocks or published illustrations of materials which are evidently welded tuffs. As in the western United States,

criteria for recognition were not known to the early workers.

Welded tuffs occur in Tertiary strata of the Boleo copper district in Lower California; the samples were collected by Wilson and studied by the writer. Immense thicknesses of welded tuffs have been recognized in western Chihuahua by W. F. Foshag and Wilfrid Kelley (personal communications). They have been studied by E. G. Zies, J. W. Greig, and Howell Williams (personal communications) in Guatemala and extend throughout Central America.

Fenner (1948) has described tuffs of Pleistocene age in the Arequipa region of southern Peru, but these are only part of the tuffs of the area. Fenner calculated that they extend northward for about 158 miles. The canyons "quebradas") cutting through these deposits are 1000 to 1500 feet deep. No estimate of the volume of these tuffs was attempted, but it must be of the order of many hundreds and probably thousands of cubic miles. Other occurrences no doubt continue the full length of the Andes.

In New Zealand Patrick Marshall (1935) has described pyroclastic materials of Pleistocene age, which he called ignimbrites; they cover an area of about 10,000 square miles and have a volume of about 200 cubic miles. Overlying the ignimbrites are great deposits of ashfall tuffs.

Richards and Bryan (1934) have described welded tuff deposits of Lower Triassic age, in the Brisbane region of southwest Queensland, Australia, which have a volume of about 43 cubic miles.

Van Bemmelen (1930; 1939) has studied the volcanic history of the Lake Toba region in northern Sumatra, and Westerveld (1947) has described the pyroclastic materials of the region as welded rhyolite tuffs of Quaternary age. These tuffs cover an area of 20,000 to 30,000 square kilometers, and Van Bemmelen (1939, p. 130) estimated that they had a volume of 2000 cubic kilometers.

Solov'ev (1950) has described large areas of welded tuffs and related pyroclastic, rhyolitic rocks from the Sikhote-alin region in easternmost Siberia, between the Amur River and the Sea of Japan. This group of rocks "originated principally during Cretaceous and lower Tertiary times" (p. 211) and covers more than 30,000 square kilometers (11,500 square miles).

Welded tuffs in Russia, Armenia (Zavaritsky, 1946), and Hungary are also well known but have been inadequately studied. They occur in Wales, France, Germany, and seemingly in Sweden, but very little is known about them in European regions.

Darwin (1843) described a "white tuffaceous rock" of Tertiary age in Patagonia as 800 feet thick, extending along the coast for 500 miles, and probably 200 miles wide. The material consists largely of fresh and altered pumice and glass, diatoms, sponge spicules, feldspar, and augite crystals. He stated that diatoms formed at least one-tenth of the deposit, the remainder being volcanic ash. The total volume thus indicated would be 15,000 cubic miles. The area may have been overestimated, but even with the most drastic reductions from the assumed figures the volume of pyroclastic material indicated is tremendous.

Dispersal of Volcanic Dust

Throughout geologic time explosive volcanic eruptions have been throwing ash into the air, and some of this received world-wide distribution. This is exemplified by Krakatoa; although its eruption was a minor episode compared with countless eruptions of the past, its effects were world-wide. A deposit of dust an inch thick and with a world-wide distribution would have a volume of a little more than 3000 cubic miles. The actual thickness is extremely problematical, but the total accumulation through the ages must be great.

The pyroclastic materials here described are in very large part Cretaceous and Tertiary. There are important occurrences that are Pleistocene in age, and the great Sumatra tuff beds are described as being Quaternary. Thus they belong for the most part to the later geologic periods. It seems improbable that explosive volcanic eruptions producing rhyolitic tuffs should be so sharply localized in geologic time. This leads one to inquire if these same materials were not produced in comparable amounts in earlier geologic time. In the older rocks may they not have been removed by erosion, concealed by overlying beds, remained

unrecognized and perhaps in part unrecognizable, as a result of metamorphic processes? Some evidence of pyroclastic materials have been observed among the older rocks in the eastern United States, even in the crystalline rocks. No doubt our Canadian friends can tell us of some in the Canadian shield. When we think how long great areas went unrecognized in rocks which had not undergone metamorphism, it is easy to believe that the crystalline rocks also contain important amounts of pyroclastic materials. The inadequate work done so far shows that pyroclastic materials, including ash beds, tuff beds, welded tuff, and bentonites in the United States alone must be measured in many tens of thousands of cubic miles.

These preliminary studies show that pyroclastic materials are present in many regions of the world, in volumes which dwarf many batholiths. An interesting comparison may be made between rhyolite pyroclastic materials and intrusive rocks of similar chemical composition—that is, with granites—but excluding the great granodiorite batholiths. This indicates that pyroclastic rhyolites occur in volumes which compare with, if they do not exceed, the volumes of true granites. It is also evident that this relationship must receive full consideration in any theories of petrogenesis.

References Cited

Allen, Victor T., 1929, Altered tuffs in the Ordovician of Minnesota: Jour. Geology, v. 37, p. 239–247.

—— 1932, Ordovician, altered volcanic material in Iowa, Wisconsin and Missouri: Jour. Geology, v. 40, p. 259–269.

Bailey, Thomas L., 1926, The Gueydan, a new Middle Tertiary formation from the southwestern Coastal Plain of Texas: Univ. Texas Bull. 2645, p. 179.

Bonine, C. A., and Honess, A. P., 1929, Bentonite in Pennsylvania: Pa. Acad. Sci. Proc., v. 3, p. 23.

Bramlette, M. N., 1946, The Monterey formation of California and the origin of its siliceous rocks: U. S. Geol. Survey Prof. Paper 212, p. 57.

Callaghan, Eugene, 1939, Volcanic sequence in the Marysvale region in southwestern-central Utah: Am. Geophys. Union Trans., 20th Ann. Meeting, p. 438–452.

Cross, Whitman, and Larsen, E. S., Jr., 1935, A brief review of the geology of the San Juan region of southwestern Colorado: U. S. Geol. Survey Bull. 843, 138 p.

Darwin, Charles, 1843, Voyage of the Beagle: N. Y., D. Appleton & Co., new ed. 1883, p. 170.

Fenner, C. N., 1948, Incandescent tuff flows of southern Peru: Geol. Soc. America Bull., v. 59, p. 879–893.

Gilbert, C. M., 1938, Welded tuff in eastern California: Geol. Soc. America Bull., v. 49, p. 1829–1862.

Griggs, Robert F., 1922, Valley of Ten Thousand Smokes: Nat. Geog. Soc., Washington, p. 29.

Hass, Wilbert, 1948, Upper Devonian bentonite in Tennessee: Am. Assoc. Petroleum Geologists Bull., v. 32, p. 816–819.

Hurley, P. M., 1951, Heat flow and chemical segregation of the mantle: Am. Geophys. Union. Prog. Abst., p. 328.

Kay, Marshall, 1931, Stratigraphy of the Ordovician Hounsfield metabentonite: Jour. Geology, v. 39, p. 361–376.

Kirkham, V. R. D., 1931, Igneous geology of southwestern Idaho: Jour. Geology, v. 39, p. 564–591.

Maddox, D. C., 1930, Bentonite in the Ordovician near Collingwood, Ontario: Science, n. ser., v. 72, p. 630.

Mansfield, G. R., and Ross, C. S., 1935, Welded tuffs in southeastern Idaho: Am. Geophys. Union Trans., 16th Ann. Meeting, p. 308–321.

Marshall, Patrick, 1935, Acid rocks of the Taupo-Rotorua volcanic region: Royal Soc. New Zealand Trans., v. 66, p. 1–44.

Moody, C. L., 1949, Mesozoic igneous rocks of northern Gulf Coastal Plain: Am. Assoc. Petroleum Geologists Bull., v. 33, p. 1410–1428.

Nelson, W. A., 1922, Volcanic ash in Tennessee, Kentucky and Alabama: Geol. Soc. America Bull., v. 33, p. 605–615.

Richards, H. C., and Bryan, W. H., 1934, The problem of the Brisbane tuff: Royal Soc. Queensland Proc., v. 45, no. 11, p. 50–62.

Ross, C. R., Miser, H. D., and Stephenson, L. W., 1930, Water-laid volcanic rocks of Early Upper Cretaceous age in southwestern Arkansas, southeastern Oklahoma and northeastern Texas: U. S. Geol. Survey Prof. Paper 154-F, p. 175–200.

Solov'ev, S. P., 1950, The principal features of the complex of young acid effusives and ignimbrites in southern Sikhote-alin, and its petrochemical characteristics: Soc. Russe Mineral., Mem., v. 79, p. 211–222.

Stephenson, L. W., 1936, Bentonite in the Upper Cretaceous of New Jersey: Science, n. ser., v. 84, p. 489.

Swineford, Ada, and Frye, J. C., 1946, Petrographic comparison of Pliocene and Pleistocene volcanic ash from western Kansas: Univ. Kans. Pub., St. Geol. Survey Kans. Bull. 64, pt. 1, p. 1–32.

Van Bemmelen, R. W., 1930, The origin of Lake Toba (North Sumatra): 4th Pac. Sci. Cong. Proc., v. IIA, p. 115–124.

—— 1939, Geologische waarnemingen in de Gajolanden (N. Sumatra): Jaarb. Mijin. in Nederl.-Indie, 59 Verhand. III, p. 71–94.

Verbeek, R. D. M., 1886, Krakatua: Batavia, p. 140.

Westerveld, J., 1942, Welded rhyolitic tuffs or ignimbrites in the Pasoemah region, West

Palembang, South Sumatra: Leidsche Geol. Mededeelingen Decl. 23, Afl. 1, p. 202–217.

Westerveld, U., 1947, On the origin of the acid volcanic rocks around Lake Toba, North Sumatra: Verhand. Konink. Nederl. Akad. Wetenschappen, Afd. Natuur. Fed. Sect., v. 43, no. 1, p. 3–51.

Williams, Howel, 1942, The geology of Crater Lake National Park, Oregon: Carnegie Inst. Washington, p. 1–151, Pl. 31.

Zavaritsky, A. N., 1946, On the Quaternary volcanic tuffs of Armenia: Acad. Nauk. Leningrad Daklady, 53, no. 8, p. 729–731.

Zirkel, Ferdinand, 1876, Geologic exploration of the Fortieth Parallel, Clarence King in charge: U. S. Army Eng. Prof. Paper 18, p. 1–175, Pl. I–XII.

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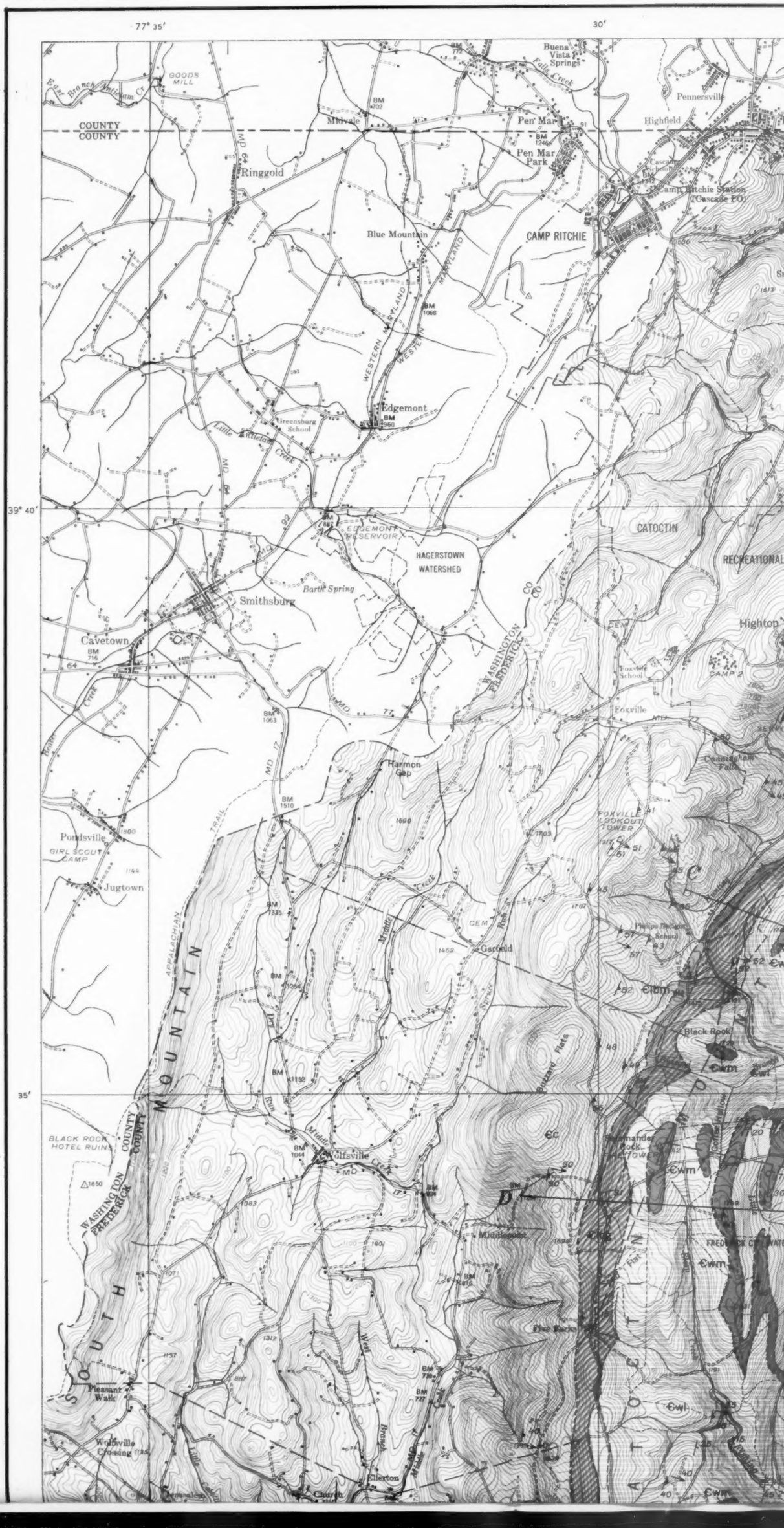
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Blue Mountain
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Greensburg School
Little Antietam Creek
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Barth Spring
Smithsburg
Cavetown
Foxville
Foxville School
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Cunningham Falls
Harmon Gap
FOXVILLE LOOKOUT TOWER
Pondsville
GIRL SCOUT CAMP
Jugtown
APPALACHIAN TRAIL
Garfield
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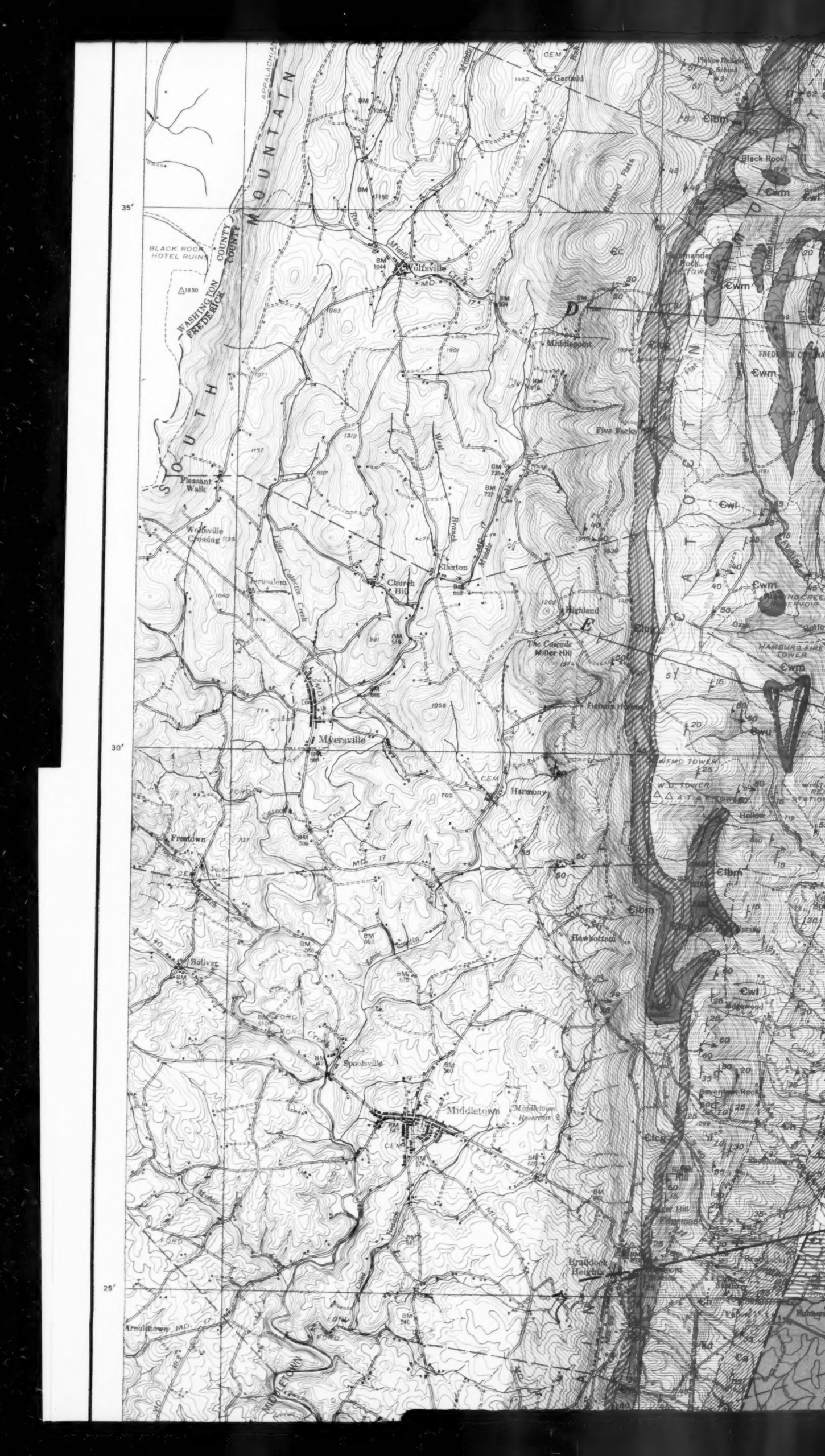
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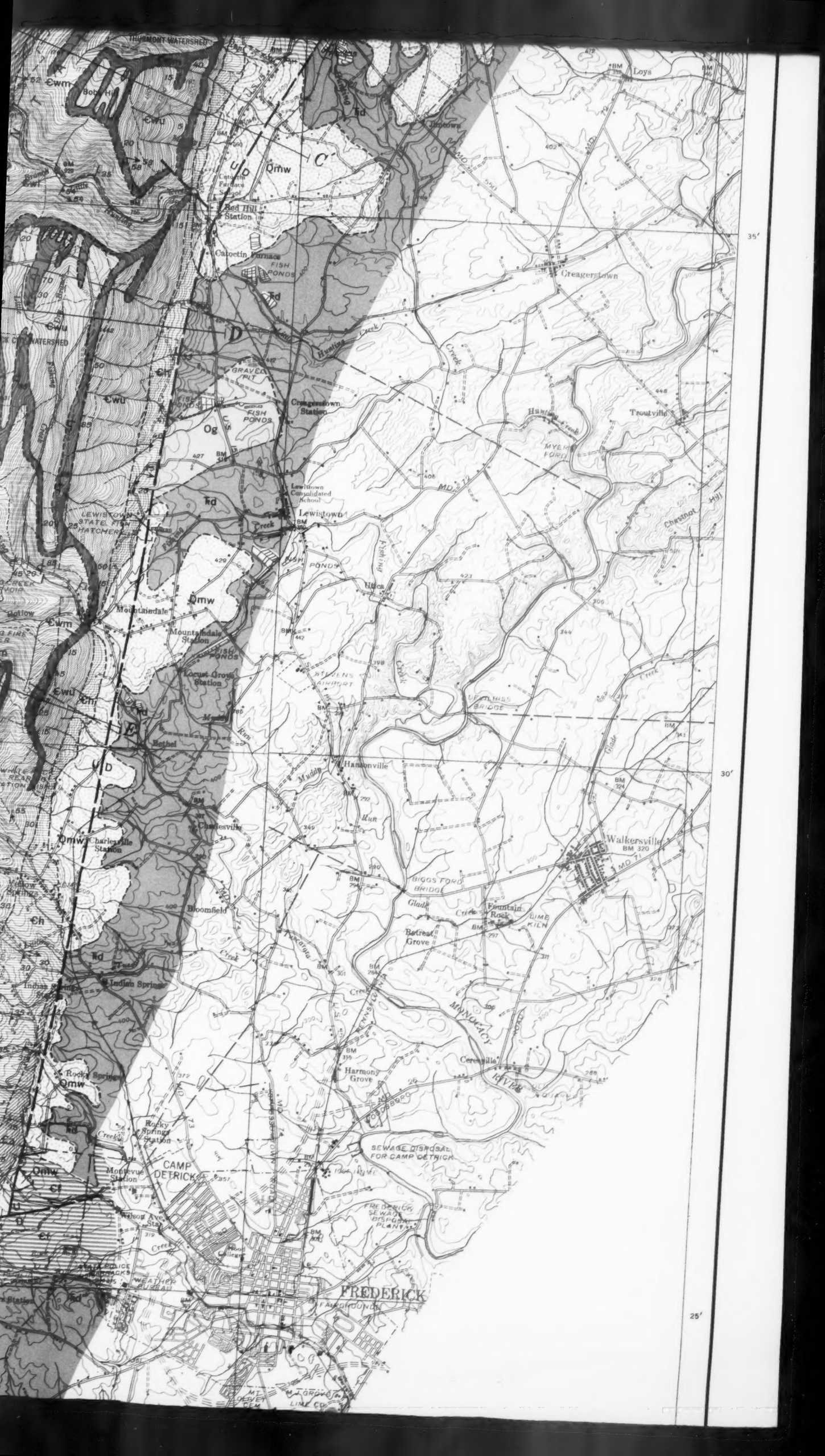
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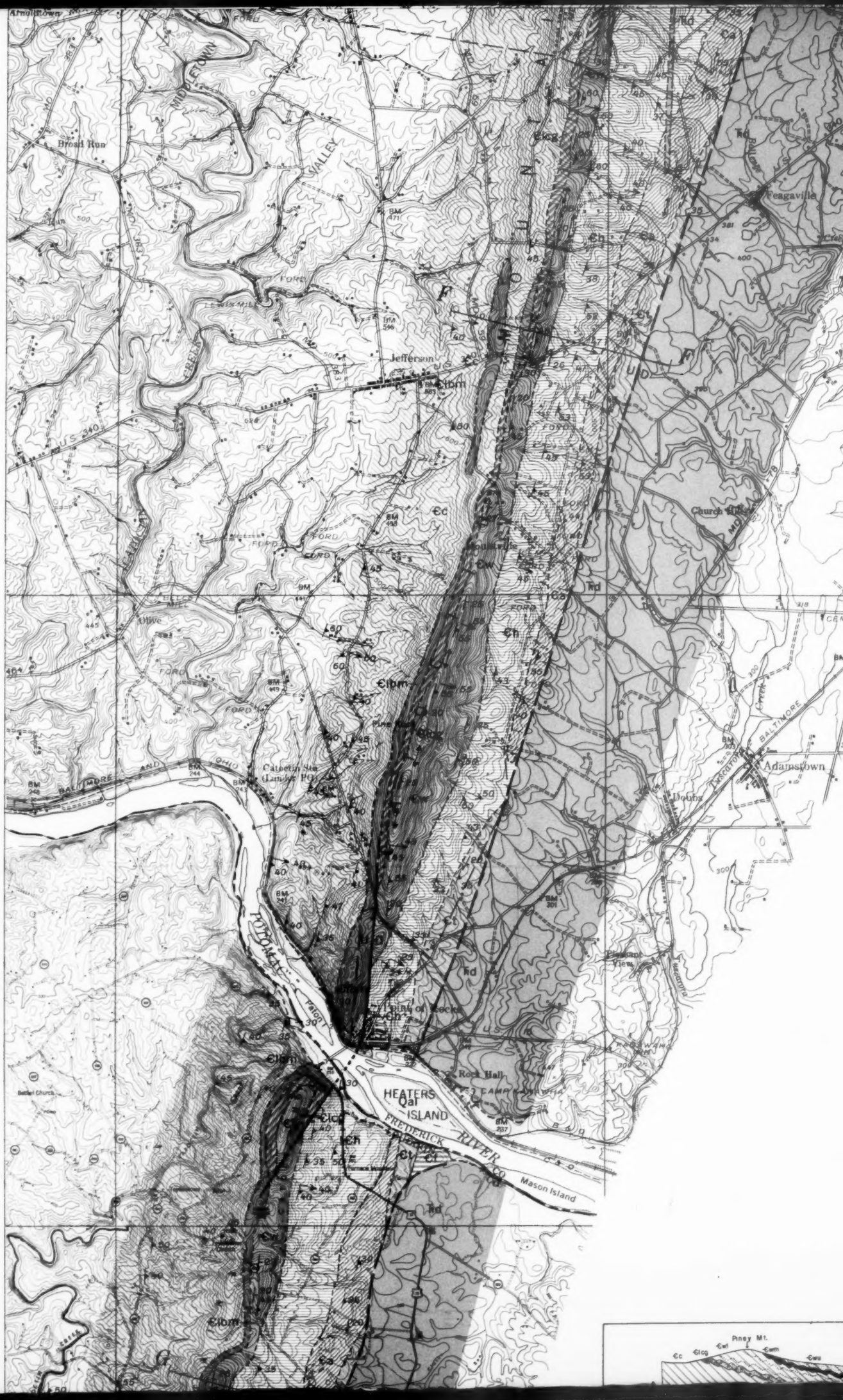
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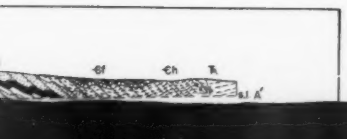
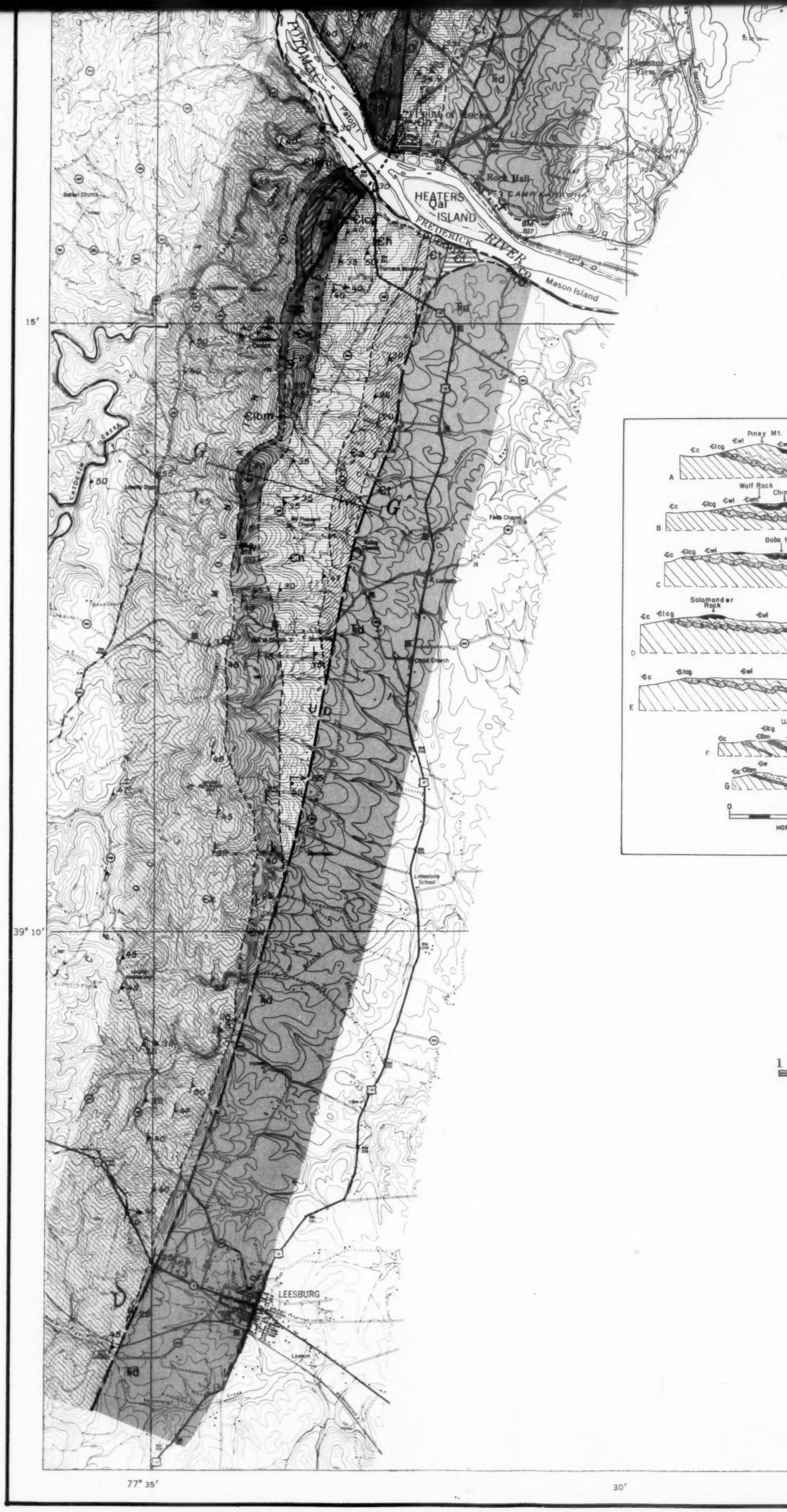
Point of Rocks
Rock Hall
HEATERS
ISLAND
FREDERICK
RIVER
Mason Island
Bethel Church
Pleasant View
Piney Mt.
20'
15'

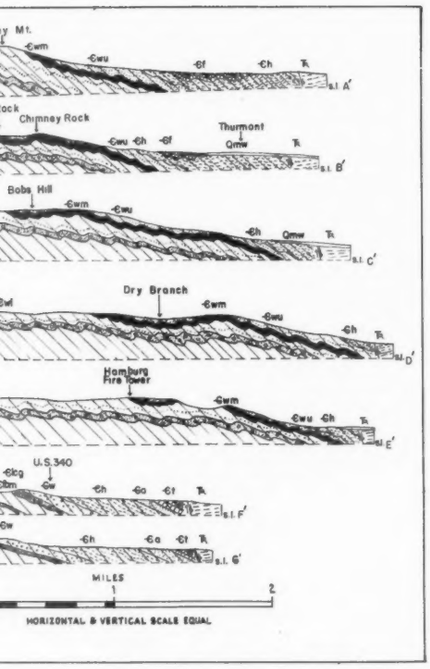
EXPLANATION

	Qal — Alluvium	QUATERNARY	
	Qmw — Alluvial mountain wash	QUATERNARY	
	Trd — Diabase and Newark series, undifferentiated	TRIASSIC	
Lower Ordovician	Og — Grove limestone	ORDOVICIAN	
Upper Cambrian ?	€f — Frederick limestone	CAMBRIAN	
	€t — Tomstown formation	CAMBRIAN	
	€a — Antietam quartzite	CAMBRIAN	
	€h — Harpers phyllite	CAMBRIAN	
Lower Cambrian	€w — €wu, €wm, €wl — Weverton quartzite *Members mapped only in northern part of area*	CAMBRIAN	
	€lcg, €lbm — Loudoun formation	CAMBRIAN	
(?)		(?)	





GEOLOGIC MAP AND PROFILES, CATOCT



Upper Cambrian(?)

Lower Cambrian

Lower Cambrian(?)

Frederick limestone

Tomstown formation

Antietam quartzite

Harpers phyllite

Weverton quartzite
Members mapped only in northern part of area

Loudoun formation

Catoctin formation

CAMBRIAN

CAMBRIAN(?)

Strike and dip of bedding

Strike and dip of overturned bedding

Strike and dip of cleavage

Horizontal projection and plunge of fold axes

Horizontal fold axes

Horizontal projection of lineation

Fault, dashed where approximately located
U, *upthrown side;* D, *downthrown side*

Scale 1:62,500

1 0 1 2 3 4 Miles

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CTIN MOUNTAIN, MARYLAND AND VIRGINIA

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 66, PP. 435-462, 9 FIGS., 3 PLS. APRIL 1955

GEOLOGY OF CATOCTIN MOUNTAIN, MARYLAND AND VIRGINIA

BY JOHN C. WHITAKER

ABSTRACT

Detailed mapping of bedding, cleavage, and lineation in the Lower Cambrian clastic rocks and underlying metavolcanics of Catoctin Mountain in Maryland and northern Virginia indicates that the area is not a syncline as previously interpreted, but a tightly folded, eastward-dipping sequence which forms the upper and eastern limb of the South Mountain anticlinorium. This interpretation necessitates a revision of the stratigraphy. Strata mapped as Loudoun along the eastern border of Catoctin Mountain constitute a zone including the upper Weverton quartzite and the basal Harpers phyllite.

Evidence suggests that the Catoctin metabasalt and associated rhyolite tuffs are closely related in time to the overlying Chilhowee group and that volcanic activity continued during deposition of the lower portion of the Chilhowee group.

A statistical study of the megascopic structures, supplemented by petrofabric analysis, contributes to an understanding of the deformation plan. Folds are asymmetrically overturned to the west; flow cleavage dips eastward and fans gently toward the west; the principal movement has been along flow-cleavage planes so that shear folds result with thickened crests and troughs and thinned limbs; fold axes are subhorizontal; lineation in *a* results from a crenulation of the flow cleavage and a stretching of minerals down the dip of the flow-cleavage plane; and the intersection of flow cleavage and bedding produces a lineation in *b*.

CONTENTS

TEXT

	Page
Introduction	436
Location of area	436
Previous work	436
Method of attack	436
Acknowledgments	436
Stratigraphy	436
General statement	436
Catoctin formation	438
General description	438
Metabasalt	438
Rhyolite tuff	439
Age of Catoctin formation and its bearing on Cambrian-Precambrian boundary	439
Loudoun formation	441
General	441
Basal phyllite member	441
Conglomerate member	442
Weverton quartzite	442
General	442
Banded quartzite	443
Thin-bedded quartzite	443
Laminated micaceous quartzite	443
Interbedded quartzite and phyllite	444
Ferruginous quartzite	444
Cross-bedded quartzite	444
Ledge-maker quartzite	445
Ferruginous quartz conglomerate	445
Harpers phyllite	445
Antietam quartzite	446
Tomstown dolomite	447
Frederick limestone	448
Stratigraphic relationship of upper Chilhowee group and overlying Cambrian carbonate rocks	448
Grove limestone	450
Triassic diabase	450
Quaternary alluvium	450
Metamorphism	451
Megascopic structures	451
Introduction	451
Primary structures	451
Secondary structures	451
Deformation plan	456
Structural interpretation	456
Introduction	456
Map pattern	457
Structure of Catoctin Mountain	457
Triassic border fault	458
Petrofabrics	458
Introduction	458
Mica orientation	460
Quartz orientation	460
Conclusions	461
References cited	461

ILLUSTRATIONS

Figure	Page
1. Index map of Blue Ridge Province of Maryland and adjacent States	437
2. South Mountain fold	438
3. Hypothetical stratigraphic relationships of upper Chilhowee group to overlying Cambrian carbonate rocks	449

Figure Page
4. Equal-area projections ... 452
5. Geologic map of area immediately east of Hamburg fire tower ... 454
6. Equal-area projections ... 455
7. Deformation plan ... 456
8. Stratigraphic relationships of Weverton ... 459
9. Petrofabric diagrams ... 460

Plate Facing page
1. Geologic map of Catoctin Mountain, Maryland and Virginia ... 465
2. Photomicrographs of Loudoun phyllite, Harpers phyllite, Tomstown dolomite, and Triassic diabase ... 466
3. Folds in Weverton quartzite ... 467

Introduction

Location of Area

Catoctin Mountain forms the easternmost portion of the Blue Ridge Province of Maryland and northern Virginia. The area mapped (Fig. 1) forms a long, narrow belt of Lower Cambrian sediments underlain by metavolcanic rocks of questionable age. The mountain extends 50 miles along the strike of the formations from Emmitsburg, Maryland, to Leesburg, Virginia. To the east lies the Cambro-Ordivician carbonate sequence and the thick Triassic sediments of the Frederick Valley, and on the west the metavolcanic rocks of the Middleton Valley. In northern Maryland the mountain is 2–4 miles wide and consists of rugged ridges rising to 1900 feet elevation. Farther south in Maryland and northern Virginia the topography is subdued. Elevations average 1000 feet, and the mountain narrows to a ridge less than 1 mile wide.

Previous Work

Keith (1892; 1893) and Stose and Stose (1946) concluded that both Catoctin and South mountains (Fig. 1) form asymmetric synclines overturned to the west.

In contrast, Cloos (1941; 1946; 1947; 1951) has shown that South Mountain is not a syncline but the overturned limb of a large anticlinorium. Further reconnaissance work led him to conclude that Catoctin Mountain forms the upper and eastern limb of this large anticlinorium. Cloos termed the entire structural unit from Catoctin Mountain westward into the Great Valley the South Mountain fold (Fig. 2).

The present study of Catoctin Mountain investigates the discrepancy between the views of Keith, Stose and Stose, and Cloos.

Method of Attack

The chief method of attack has been mapping on whatever scale was necessary to understand the problem. Because the folding of the quartzites is tight and complex, much of the area was mapped on the scale of 1:5000 with the aid of areal photographs and stereoscope. Quartzite beds were traced along the strike but were plotted on the map only where they crop out or are clearly expressed topographically. A three-dimensional plaster model was constructed on the scale of 1:25,000 so that the structure and outcrop pattern of the formations could be readily visualized. About 4000 measurements of bedding, flow and fracture cleavage, lineations, fold axes, and joints were made. These were plotted in the lower half of an equal-area projection in order to obtain statistical maxima; only representative values have been plotted on the map (Pl. 1).

Petrofabric diagrams have been prepared, and 53 thin sections studied.

Acknowledgments

The writer is indebted to Ernst Cloos, who suggested the problem and gave freely of his time for instruction and criticism. The author also wishes to express his gratitude to Thomas W. Amsden, who made many valuable suggestions concerning the stratigraphy of the area, and to Aaron C. Waters, whose critical reading of the manuscript was of great assistance. The guidance of the Reverend Gerard Yates, S.J., and Josef Solterer is deeply appreciated.

Stratigraphy

General Statement

The core of the South Mountain uplift consists of Precambrian granodiorite and

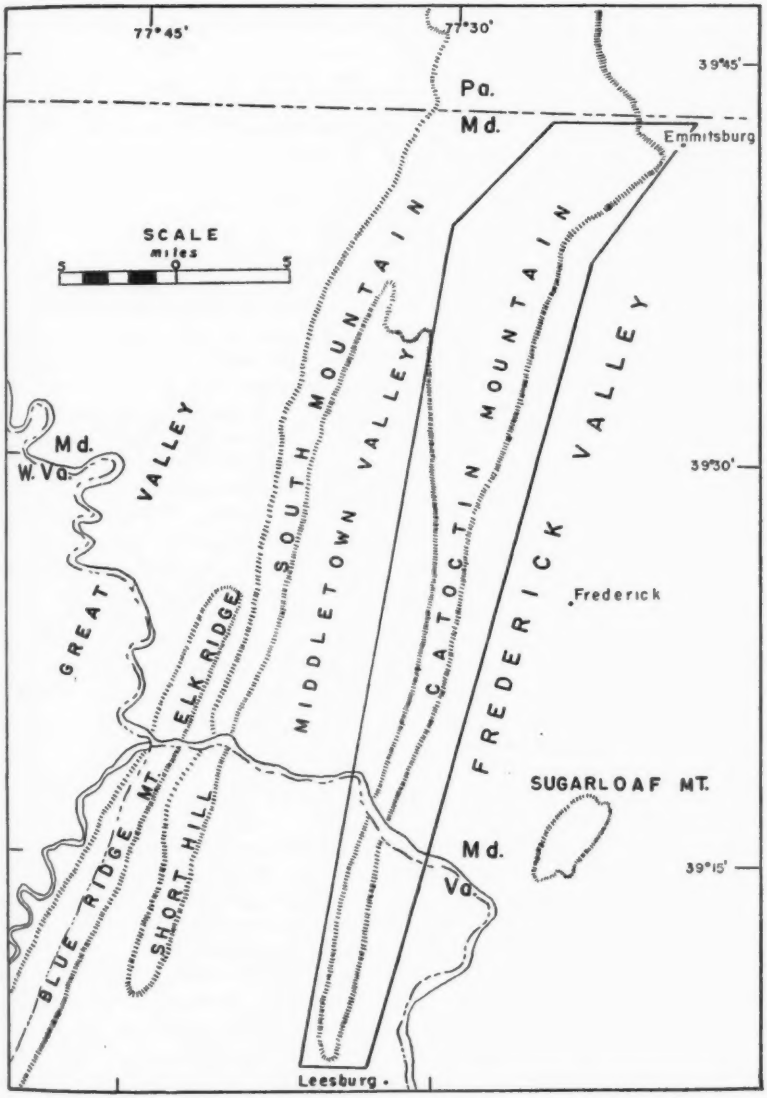

FIGURE 1.—INDEX MAP OF BLUE RIDGE PROVINCE OF MARYLAND AND ADJACENT STATES

biotite granite gneiss. Overlying this are thin, lenticular bodies of tuffaceous phyllite and thin-bedded, micaceous quartzite which have been named both the Swift Run formation (Jonas and Stose, 1939, p. 583) and the Oronoco formation (Bloomer and Bloomer, 1947, p. 95) from localities in northern Virginia. Stose and Stose (1946, p. 18) have traced this same lithology northward into Maryland.

Overlying the Catoctin metavolcanic rocks is about 3000 feet of clastics termed the Chilhowee group (Safford, 1856, p. 152–153). The

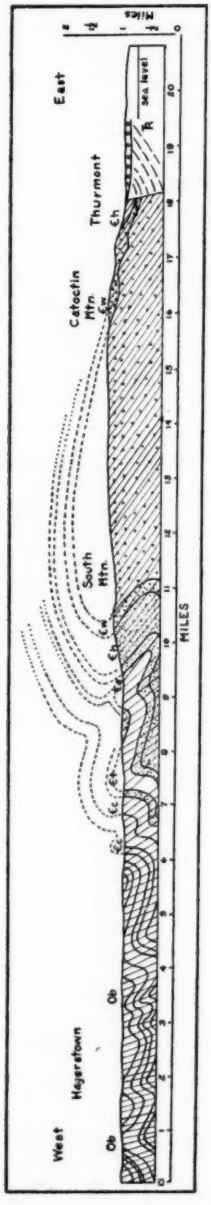

FIGURE 2.—SOUTH MOUNTAIN FOLD
After Cloos (1947, p. 857)

Chilhowee group is one of the most widespread units of the southern Appalachians. Representatives of the group occur in Tennessee, Virginia, and Maryland; equivalents extend southwest into Alabama and northeast beyond Pennsylvania (King, 1949, p. 519). These sediments form Catoctin, South, and Blue Ridge mountains as well as Short Hill and Elk Ridge (Fig. 1). Most writers consider the Chilhowee group lower Cambrian; the entire group forms a conformable sequence, and the upper portion contains diagnostic lower Cambrian fossils. In Maryland and northern Virginia, the Chilhowee group consists of the Loudoun phyllite and conglomerate, the Weverton quartzite, the Harpers phyllite, and the fossiliferous Antietam quartzite.

Overlying the Chilhowee clastics is the carbonate sequence which underlies both the Frederick and Great valleys. In the Frederick Valley, the carbonate sequence is partially covered by Triassic sediments so that only the Cambrian Tomstown dolomite, Frederick limestone, and the overlying Grove limestone of early Ordovician age have been recognized along the east side of Catoctin Mountain.

The basement complex and the overlying Swift Run lie outside the immediate area mapped and are not described.

Catoctin Formation

General description.—The Catoctin formation was first named the Catoctin schist (Geiger and Keith, 1891, Pl. 4) after exposures along the western border of Catoctin Mountain. In Frederick County, Maryland, Stose and Stose (1946, p. 20) have recognized metabasalt, andesite, aporhyolite, and rhyolite tuff. In the immediate vicinity of Catoctin Mountain, the writer recognized the dense, hard, moderate-green (5G5/6) (Goddard *et al.*, 1951) rock termed metabasalt by Stose and Stose as well as the pale-purple (5P6/2) phyllite or rhyolite tuff.

Metabasalt.—The metabasalt is a moderate-green schistose rock, in many places amygdaloidal, strongly lineated, and intersected by numerous quartz and epidote veins. The lineation is accentuated by blebs of chlorite elongated down the dip of the flow-cleavage plane,

ard by stretched amygdules filled with quartz, calcite, and epidote.

In thin section, the metabasalt consists of chlorite, epidote, and albite laths with smaller amounts of actinolite, magnetite, and ilmenite. No glass is present, and most of the minerals have been replaced. Chlorite and actinolite replace pyroxene, and epidote replaces albite. An ophitic network of albite laths with interstitial chlorite and epidote is common.

The epidote is probably a manganiferous variety. Under crossed nicols it is deep yellow to brown. The measured indices are: $N\alpha = 1.722$, $N\beta = 1.743$, and $N\gamma = 1.758$. Piedmontite, the red, manganese-bearing epidote, has been noted in vesicles (Stose and Stose, 1946, p. 20).

The writer would justify the term metabasalt. Five Rosiwal analyses by the writer, in addition to the chemical analysis by Henderson (Stose and Stose, 1946, p. 21) indicate a composition similar to that for the average unaltered basalt (Barth, 1952, p. 69). Also, the ophitic network of feldspar laths with interstitial epidote and chlorite indicates the texture of a normal basalt.

Rhyolite tuff.—Pale-purple tuffaceous phyllite occurs as float along the eastern border of the Middletown Valley. The stratigraphic relationship of these phyllites to the metabasalt is in doubt. Stose and Jonas (1938) mapped two bodies of pale-purple tuffaceous phyllite in the Middletown Valley south of Jefferson, Maryland, and considered them equivalent to the phyllites of the Loudoun formation at Catoctin Mountain; hence they interpret these tuffaceous phyllites as infolds of Loudoun in the underlying metabasalt. They map the two bodies of phyllite as synclines, both connecting at their southern end with the phyllite of the Loudoun at Catoctin Mountain. The writer mapped one of these bodies of tuffaceous phyllite south of Jefferson, Maryland, but found no connection between it and the phyllites of the Loudoun at Catoctin Mountain. The best exposure of the stratigraphic relationship of the tuffaceous phyllite and the metabasalt is outside the area in a road cut on Pennsylvania State Highway 16, 0.2 mile east of the intersection of Pennsylvania State Highway 81 east of Pen Mar, Pennsylvania. The road cut exposes the gently dipping east limb and nearly vertical west limb of an anticline of tuffaceous phyllite with angular fragments of metabasalt, quartz, and feldspar. Well-rounded grains of blue quartz are also present. The tuffaceous phyllite grades into metabasalt both above and below. Thus the purple tuffaceous phyllite is probably interbedded with metabasalt flows throughout much of the Middletown Valley.

Both Stose and Stose (1946, p. 20) and Bascom (1896) have called the tuffaceous phyllite a rhyolite tuff. Undeformed specimens collected farther north in Franklin County, Pennsylvania, show this very clearly (Bascom, 1896). Thin sections from the metamorphosed phyllite in the vicinty of Catoctin Mountain consist of sericite and angular fragments of quartz, feldspar, and epidote. Abundant magnetite as well as ilmenite altered to leucoxene are present. The feldspathic constituents and original glass shards have altered to sericite, but the outlines of some of the original angular fragments of feldspar and quartz remain.

Age of Catoctin formation and its bearing on Cambrian-Precambrian boundary.—The position of the base of the Cambrian in the Southern Appalachians is much debated. The Committee on Stratigraphy of the National Research Council (Howell *et al.*, 1944) places the base of the Cambrian at the lowest formation containing diagnostic fossils (the Antietam in Maryland). Snyder (1947, p. 152) and Wheeler (1947, p. 157–159) agree with this procedure, pointing out that since the limits of all the other systems are defined paleontologically, the Cambrian should be no exception.

Since several thousand feet of sediments is conformable below the lowest diagnostic Cambrian fossils of the Chilhowee group, most writers have attempted to draw the base of the Cambrian at some significant hiatus below the lowest fossiliferous beds. This procedure emphasizes the close time relation of the unfossiliferous sediments to the overlying fossiliferous Cambrian strata. At the same time, it raises disturbing problems. How profound an unconformity must be present to justify selecting the base of the Cambrian? Indeed, how does one tell a short from a long hiatus without the aid of fossils? And finally, unconformities do not necessarily represent time lines.

The problem of the base of the Cambrian in the Southern Appalachians has been thoroughly reviewed by King (1949). The writer therefore restricts his comments to the Maryland and northern Virginia area in the hope that they will be a contribution to this regional problem.

Stose and Stose (1946, p. 28) indicate an unconformity between the base of the Chilhowee (Loudoun) and the underlying Catoctin metabasalt and therefore place the base of the Cambrian between the two. They state that the Loudoun overlaps on Swift Run, aporhyolite, metabasalt, and the basement complex. They also call attention to diabase dikes genetically related to the metabasalt which intrude the basement complex but not the overlying Chilhowee. Moreover, the writer found extensive conglomerates at the base of the Chilhowee, at Catoctin Mountain, and Cloos (1951, p. 27–29) has described conglomerates locally developed in a corresponding stratigraphic position at South Mountain.

In contrast, Cloos (1951, p. 27) suggested that the Catoctin formation might be Cambrian because there is (1) structural conformity between the volcanics and the overlying Chilhowee sediments, (2) gradational contact between sandy volcanics and tuffaceous sediments, and (3) no clear-cut unconformity between the two.

The writer feels there is no definite evidence for an unconformity between the metabasalt and the overlying Chilhowee sediments and that the only significant hiatus occurs between the basement complex and the overlying Swift Run formation. The reasons are as follows:

(1) Purple tuffaceous phyllites are interbedded with the metabasalt and grade upward into the Chilhowee group as high as the basal portion of the Harpers. Tuffaceous phyllites, similar to those interbedded with the metabasalt described above, have been mapped as lenticular, discontinuous bodies that form the basal member of the Loudoun (Pl. 1). These same tuffaceous phyllites form the matrix of the overlying Loudoun conglomerate. Purple tuffaceous phyllites occur interbedded with the Weverton and continue upward with decreasing frequency into the basal Harpers. It therefore seems reasonable that volcanic activity commenced with the Swift Run formation, reached its culmination as metabasalt flows and interbedded tuffs, and waned during Chilhowee time. The inference is that this represents one cycle of volcanic activity with no recognizable hiatus.

(2) Blue quartz, so characteristic of the Chilhowee group, also occurs with the tuffs associated with the metabasalt flows, suggesting the same source for both the Chilhowee sediments and the blue quartz interbedded with the volcanics. Thus it appears that the Chilhowee-type sediments were being supplied even during the culmination of volcanic activity and that the same source area may have supplied the sediments from the initial phase of volcanic activity until the end of Chilhowee time.

(3) The chief point in favor of placing the base of the Cambrian between the Chilhowee and the underlying volcanics is that the Loudoun conglomerate occurs almost continuously along the strike of Catoctin Mountain near the base of the Chilhowee group. These conglomerates, however, contain no fragments of metabasalt; if an unconformity exists here it is reasonable to expect metabasalt fragments derived from the erosion surface of the volcanics.

(4) The basal tuffaceous phyllite member of the Loudoun has been mapped as irregular lenses pinching out along the strike between the Loudoun conglomerate and the underlying metabasalt. This suggests irregular deposition on a volcanic surface of comparatively rugged topography with the tuff occupying the topographic basins. This condition would result whether the tuff be water laid or due to ash falls. Keith (1893, p. 327–329) thought the pre-Loudoun surface of the Catoctin metabasalt in northern Virginia probably had a relief of 500–800 feet. Stose and Stose (1946, p. 31–34) suggest that the pre-Loudoun surface in Maryland was more perfectly peneplaned than it was considered to be by Keith in northern Virginia. The writer grants that such a surface of relief may represent a long period of erosion. On the other hand, an area of volcanic activity might be expected to have considerable relief without being subjected to erosion of any considerable duration. The relief of the pre-Loudoun surface, therefore, may or may not have any appreciable time significance.

(5) Similarily, the overlap described by Stose and Stose (1946, p. 28) may be due merely to an irregular distribution of the volcanics above the basement complex. There is no evidence that this overlap means that the volcanics were folded and eroded before the Chilhowee was deposited.

(6) The fact that diabase dikes genetically related to the volcanics intrude the granodiorite and not the overlying Chilhowee group merely means that the clastic sediments had not yet been deposited (Cloos, 1951, p. 27).

The writer, therefore, concludes that volcanic activity began after prolonged erosion of the basement complex, climaxed in the Catoctin metabasalt flows and interbedded tuffs, and waned during Chilhowee time. Thus, there seems to be one phase of volcanic activity with no recognizable hiatus; the sediments were deposited on a gradually subsiding volcanic surface. Whether the Catoctin formation will be classified as Cambrian or Precambrian remains a philosophical question. It is more important that the volcanics seem to be closely conformable and closely related in time to the overlying Chilhowee sediments, whose upper portion contains Lower Cambrian fossils.

Loudoun Formation

General.—The name Loudoun formation was given (Keith, 1893, p. 324; Williams and Clark, 1893, p. 68) to strata exposed along Catoctin Mountain in Loudoun County, Virginia. No fossils have been found in the Loudoun, but it is here considered Lower Cambrian because of its structural and stratigraphic relationship to the overlying Antietam quartzite.

The stratigraphy of the Loudoun needs revision. Keith (1893, p. 324) described it as a heterogeneous formation of conglomerate, feldspathic sandstone, purple slates, and marble, ranging in thickness from 800 feet to less than 10 feet over very short distances. Stose and Stose (1946, p. 32) describe the formation as a soft arkosic quartzite with beds of purer, harder quartzite, quartzose conglomerate, and phyllite or slate.

The writer considers both the original description of the Loudoun by Keith and the revised description by Stose and Stose invalid because:

(1) The white crystalline marble and purple slates described by Keith in the southern extension of the Middletown Valley into Virginia are not infolds of the Loudoun in the underlying metabasalt, but occur between the metabasalt and the basement complex and correspond in their stratigraphic position to the Swift Run formation (Jonas and Stose, 1939, p. 583) or the Oronoco formation (Bloomer and Bloomer, 1947, p. 95).

(2) The pale-purple tuffaceous phyllites mapped by Stose and Jonas (1938) as infolds of the Loudoun in the underlying metabasalt south of Jefferson, Maryland, appear to be tuffaceous phyllites interbedded with the metabasalt flows and therefore do not overlie the metabasalt in the stratigraphic position of the Loudoun.

(3) Both Keith (1893) and Stose and Stose (1946) interpret Catoctin Mountain as a syncline overturned asymmetrically to the west. By this interpretation the Loudoun is exposed on both the eastern and western slopes of Catoctin Mountain and encloses the overlying Weverton quartzite. The writer presents evidence that this interpretation is incorrect and that Catoctin Mountain forms an east-dipping, tightly folded sequence of Chilhowee clastics which overlies the Catoctin volcanics in normal succession. According to the writer's interpretation the Loudoun is exposed only along the western slopes of Catoctin Mountain where the formation overlies the metabasalt in normal sequence. Furthermore, the strata mapped as Loudoun on the eastern slopes of Catoctin Mountain actually form a zone including the basal portion of the Harpers and the uppermost beds of the Weverton quartzite.

The writer has mapped two members within the Loudoun formation and has assigned to them descriptive rock terms rather than geographic names. Both members are exposed along the strike for more than 40 miles, but at no place is there a complete exposure, so that no specific geographic type locality has been assigned. These two mappable units are termed the basal phyllite member and the overlying conglomerate member.

Basal phyllite member.—Along the western slopes of Catoctin Mountain, lenticular bodies of pale-purple (5P6/2) phyllite occur intermittently along the strike from the northern

portion of the area as far south as Bald Mountain in Loudoun County, Virginia. Exposures are conspicuous immediately east of the National Park Service office building in a road cut on the north side of State Highway 77, west of Thurmont. No bedding can be seen, but a prominent flow cleavage strikes N.25°E. and dips 65°E. On the steep cliffs around High Knob in Gambrill State Park the basal phyllite is well exposed where it overlies the metabasalt. In southern Frederick County and northern Loudoun County the basal phyllite member occurs only as float scattered along the western slope of Catoctin Mountain. The thickness of the basal phyllite member is variable with a maximum of about 100 feet.

In thin section, the basal phyllite member of the Loudoun is similar to the rhyolite tuff interbedded with the metabasalt flows. The rock consists of angular fragments of quartz and orthoclase with a high proportion of magnetite set in a groundmass of sericite and minor amounts of chlorite. The phyllite is strongly sheared and transected by fracture cleavage (Pl. 2, fig. 1). The high proportion of sericite suggests that originally the rock was a tuff containing glass shards. An alternate view, that the rock is a metamorphosed graywacke, is unlikely because of the angularity of the fragments, the small proportion of chlorite, and the lack of biotite which would be the normal recrystallization products of the iron- and magnesium-rich clays associated with unmetamorphosed graywackes. The purple phyllites may be original ash falls or tuffaceous sediments transported a relatively short distance. The fact that the basal phyllite member is mapped as discontinuous, lensing bodies of variable thickness also suggests that it may have been deposited in basins on the relatively rough volcanic surface of the metabasalt.

Conglomerate member.—Overlying the basal phyllite is the conglomerate member of the Loudoun, exposed almost continuously along the western slope of Catoctin Mountain from the northern limit of the area to about 1 mile south of the Potomac River in Loudoun County, Virginia. The rock is composed of rounded quartz, feldspar, and granitic pebbles half a cm to 10 cm in size. The matrix consists of the same pale-purple phyllite as the basal Loudoun member, and the two members are in gradational contact.

In thin section the conglomerate contains strained quartz grains with inclusions of biotite and zircon. The feldspars are orthoclase and microperthite. The granitic pebbles are composed of grains of quartz, orthoclase, microperthite, and microcline with sutured boundaries. The matrix consists of angular fragments of quartz and orthoclase with a large proportion of magnetite, set in a matrix of sericite and minor amounts of chlorite.

Partial sections measured on the west side of Catoctin Mountain suggest that the conglomerate is about 100 feet thick in the northern portion of the area. At the western end of the Owens Creek gorge, on the southern side of the creek, 93 feet of the conglomerate member interbedded with thin beds of pale-purple phyllite is exposed. Immediately north of U. S. Highway 40*A*, about 200 yards east of the summit of Catoctin Mountain, 42 feet of the conglomerate member has been measured with the base of the section unexposed. On the north side of U. S. Highway 340, at the summit of Catoctin Mountain, 24 feet of the conglomerate member is exposed, but the contact with the basal phyllite is unexposed. Southward from U. S. 340, along the summit and western slopes of the mountain, ledges of the conglomerate member have been measured, but they do not exceed 30 feet in thickness. Because the thickest sections of the conglomerate member have been measured in the northern portion of the area, and because the outcrop width is considerably narrower in the southern portion of the area, the conglomerate member probably thins southward to where it pinches out about 1 mile south of the Potomac River.

Weverton Quartzite

General.—The Weverton quartzite was first differentiated, but not named, by Keith (1892, p. 365). The formation was named in separate publications by Keith (1893, p. 329) and by Williams and Clark (1893, p. 68) from exposures in the Potomac River gorge near the town of Weverton, Maryland. No fossils have been found in the Weverton, but the formation

is here considered Lower Cambrian because of its structural and stratigraphic relationship to the Antietam quartzite.

The Weverton quartzite forms the eastern slope of Catoctin Mountain from the northern limit of the area near Mount St. Marys, Maryland, to the southern limit of the area near Leesburg, Virginia. In the northern portion of the area the Weverton is thicker and more tightly folded so that the outcrop width of the formation increases with a corresponding increase in the width of the mountain. Southward the Weverton thins and forms a uniformly eastward-dipping sequence so that the outcrop width of the Weverton decreases with a similar decrease in the width of the mountain.

The Weverton forms a complex unit of clastics which vary along the strike over a fairly limited geographic area. These rock types are: banded quartzite, thin-bedded quartzite, laminated micaceous quartzite, interbedded quartzite and phyllite, ferruginous quartzite, cross-bedded quartzite, ledge-maker quartzite, and ferruginous quartz conglomerate. Figure 8 represents the facies changes which take place within the Weverton and also portrays the stratigraphic relationship of the Weverton to both the overlying Harpers and underlying Loudoun and Catoctin formations.

The most striking feature of the Weverton is the northward thickening, accompanied by an increasing complexity of lithologic units. The section thickens from approximately 100 feet in the southern portion of the area to about 350–425 feet in the northern portion. The facies changes presented in Figure 8 are generalized in that the contact between the rock types is very gradational, and as a general rule it is impossible to tell which lithology is thickening or thinning at the expense of another lithology.

Banded quartzite.—In the southern portion of the area from Leesburg to just south of Bald Mountain in Loudoun County, Virginia, a fine-grained, thinly banded quartzite overlies the Catoctin metabasalt. The total thickness is not exposed since the upper portion of the section is cut off by the Triassic border fault.

Along Limestone Branch 1⅘ miles southwest of Limestone School on U. S. Highway 15, a nearly continuous section of the fine-grained, banded quartzite is exposed. The base is not exposed, but metabasalt float occurs just beyond the western limit of the section. Since neither the basal phyllite nor the conglomerate members of the Loudoun are present, fine-grained, thinly banded quartzite appears to overlie the Catoctin metabasalt. The section from top to base is:

Feet	
98	Very light-gray (N8), hard, vitreous, fine-grained, banded quartzite. Bands composed of concentrations of blue and red quartz layers averaging 5 mm in width, spaced 2–5 inches apart
7	Concealed
7	Very light-gray, hard, vitreous, fine-grained, banded quartzite
3	Concealed
4	Very light-gray, hard, vitreous, fine-grained, banded quartzite
119	Total

Thin-bedded quartzite.—From just south of Bald Mountain in Loudoun County, Virginia, to about 3 miles north of the Potomac River in Frederick County, Maryland, the metabasalt is overlain by a thin-bedded, very light-gray (N8), vitreous quartzite similar to those just described except that the bands of concentrated blue and red quartz are absent. No complete section is exposed but the quartzite may be seen by walking along the summit of Catoctin Mountain where a thin-bedded quartzite overlies the Loudoun conglomerate. The estimated thickness from scattered exposures is approximately 100 feet.

Laminated micaceous quartzite.—From about 4 miles south of U. S. Highway 340 to just south of U. S. Highway 40*A*, a laminated micaceous quartzite makes up the Weverton formation. A complete section is exposed in the road cut of U. S. Highway 340 at the summit of Catoctin Mountain. For a distance of 165 feet west of the last white guardpost, on the north side of the highway at the eastern entrance to the road cut, pale-purple phyllite is interbedded with thin beds of micaceous quartzite. This is the pale-purple phyllite at the base of the Harpers in gradational contact with the Weverton. West of this is an almost continuous

exposure of finitely laminated, micaceous quartzite. The section, from top to base, is as follows:

Feet	
4	Finely laminated, micaceous, medium-grained, light-greenish-gray (5GY8/1) quartzite
12	Partly concealed weathered pale-blue (5PB7/2) phyllite
12	Finely laminated, micaceous, medium-grained, light-greenish-gray quartzite
28	Alternating light-greenish-gray and medium-bluish-gray (5B5/1), medium-grained, finely laminated, micaceous quartzite. Color difference due to concentration of white and blue quartz grains. Vein quartz lenses 2–10 inches long occur along bedding
10	Coarse-grained, light-greenish-gray quartzite with quartz grains averaging 3 mm
3	Pale-purple (5P6/2) phyllite
12	Coarse-grained, medium-bluish-gray quartzite with quartz grains averaging 3 mm interbedded with pale-purple phyllite beds 1–4 inches thick
6	Coarse-grained, light-greenish-gray quartzite with quartz grains as large as 5 mm
87	Total

In general, a very finely laminated, micaceous quartzite occurs near the top, and becomes more coarse-grained, thick-bedded, and conglomeratic near the base. The lowest beds grade into the Loudoun conglomerate, which is exposed just north of the west end of the road cut.

In thin section, the finely laminated, micaceous quartzite consists of large grains of strained quartz, showing sutured boundaries and undulatory extinction, in a groundmass of smaller, interlocking, anhedral grains of quartz. The micaceous laminae are predominantly sericite with smaller amounts of chlorite. A few subrounded detrital grains of pyroxene are also present.

Interbedded quartzite and phyllite.—Northward from U. S. Highway 40, much of the Weverton is composed of thin-bedded, very light-gray (N8) quartzite interbedded with dusky-yellow (5Y6/4) phyllite. Northward from U. S. Highway 40*A*, this lithology interfingers with ferruginous quartz conglomerate, ferruginous quartzite, cross-bedded quartzite, and the ledge-maker member (Fig. 8). It grades into the overlying Harpers by an increase in the proportion of phyllite near the top of the section.

Ferruginous quartzite.—In a restricted area north of U. S. Highway 40*A*, and south of Lewistown, Maryland, the Loudoun conglomerate is overlain by a poorly exposed unit of thin, cross-bedded, ferruginous quartzite. The unit is best exposed half a mile east of High Knob where a fire trail leads eastward from the paved road leading to High Knob. Exposures are not continuous, but the unit is estimated to be 75 feet thick. The rock is a fine-grained, light-gray (N7) quartzite, with thin, cross-bedded, iron-rich laminae. The laminae are composed of magnetite largely altered to hematite.

In thin section the rock is composed of quartz grains averaging 1 mm in a groundmass of fine grains of quartz and sericite. A few well-rounded, detrital grains of pyroxene and hornblende are present.

Cross-bedded quartzite.—Overlying the Loudoun conglomerate member from the northern limit of the area southward to the vicinity of Lewistown, Maryland, is a distinctive greenish-gray (5GY6/1), cross-bedded quartzite. The unit grades downward into the Loudoun conglomerate with an increase in grain size and an addition of tuffaceous phyllite, and upward into the ledge-maker member with a reduction in the amount of clayey constituent. Measurements of scattered ledges exposed in the northern portion of the area indicate that the unit is 150–175 feet thick. A partial section is readily viewed immediately east of the dirt road which follows the west branch of Fishing Creek a tenth of a mile northwest of the confluence of the east and west branches of Fishing Creek. The section has been measured from top to base in the steep cliffs immediately east of the dirt road.

Feet	
4	Fine-grained, cross-bedded, greenish-gray quartzite interbedded with pale-purple (5P6/2) phyllite beds from 1/16 to 1/8 inch wide
8	Concealed
21	Fine-grained, greenish-gray quartzite with cross-bedded dusky-yellow (5Y6/2) silty partings about 1/16 inch wide. This sequence is interbedded with pale-purple phyllite beds 2–4 inches thick

Feet	
2	Concealed
36	Medium-grained, greenish-gray quartzite with cross-bedded dusky-yellow silty partings with rhythmically repeated pale-purple phyllite beds 2–4 inches thick occurring at intervals of 2–4 feet
19	Coarse-grained, greenish-gray quartzite with cross-bedded dusky-yellow silty partings interbedded with pale-purple phyllite beds 4–7 inches thick
90	Total

Ledge-maker quartzite.—The ledge-maker quartzite member is the only well-defined mappable unit within the Weverton ("Cwm" in Pl. 1). So that the structure of the area may be more readily visualized, all the Weverton beds below the ledge-maker member have been designated as "Cwl", and those beds above as "Cwu" (Pl. 1).

The exact thickness of the ledge-maker member is unknown, but it is estimated to have a maximum thickness of 110 feet in the northern portion of the area, lensing out southward near Yellow Springs, Maryland.

In cliffs just north of Fishing Creek road 1.4 miles west of Mountaindale, Maryland, the ledge-maker member dips eastward under the Harpers phyllite. The section from the top of the cliffs downward includes part of the overlying ferruginous quartz conglomerate, but the entire ledge-maker member is not exposed:

Feet	
9	Ferruginous quartz conglomerate. White and blue rounded quartz grains up to 8 mm (average 3 mm). Bedding determined by thin ($\frac{1}{16}$ inch), iron-rich layers spaced 1–3 inches apart.
6	Very light-gray (N8), medium-grained quartzite. Bedding interval 2–6 inches, distinguished by thin ($\frac{1}{16}$ inch), dusky-yellow (5Y6/4) shaly partings.
7	Partially concealed dusky-yellow sandy phyllite
43	Massive, well-jointed, hard, very light-gray quartzite. Bedding distinguished by thin ($\frac{1}{16}$ inch), dusky-yellow shaly partings or by layers of alternating blue and white quartz grains.
10	Very massive, well-jointed, hard, very light-gray quartzite where bedding can be distinguished as very thin ($\frac{1}{32}$ inch) dark laminae
66	Total (exclusive of ferruginous quartz conglomerate)

In thin section the rock is almost a pure quartzite (orthoquartzite). The large quartz grains average 1 mm and are set in a groundmass of interlocking anhedral grains of fine quartz. The larger quartz grains show undulatory extinction and Boehm lamellae. A small amount of sericite is aligned parallel to the trace of the cleavage plane. A few rounded detrital epidote grains occupy the interstices.

Ferruginous quartz conglomerate.—A ferruginous quartz conglomerate occurs immediately below the interbedded quartzite and phyllite near the top of the Weverton from just south of Thurmont to immediately south of U. S. Highway 40*A*. This same lithology occurs from about 1 mile south of U. S. Highway 340 to about 1 mile south of the Potomac River. No complete section is exposed, but the unit is estimated to range between 15 and 40 feet. Exposures can readily be seen immediately north of the picnic area on U. S. Highway 40*A*. The rock is medium bluish-gray (5B5/1) on fresh surfaces but weathers to a mottled grayish-pink (5R8/2) and dark reddish-brown (10R3/4) with the alteration of magnetite to hematite. Many blue and white rounded quartz grains as large as 8 mm (average 2 mm) are elongated parallel to the trace of the cleavage. Bedding is made conspicuous by thin ($\frac{1}{16}$ inch) magnetite-rich layers partially altered to hematite in weathered exposures.

Harpers Phyllite

The Harpers phyllite was first differentiated but not named by Keith (1892, p. 365). It was named in separate publications by Keith (1893, p. 333) and by Williams and Clark (1893, p. 68) for exposures in the Potomac River gorge near Harpers Ferry, West Virginia. In 1896 Walcott (p. 25) recorded the supposed worm boring *Scolithes linearis* Haldemann (Stose, 1906, p. 205; Bassler, 1919, p. 58; Amsden, 1951, p. 102). The Harpers is herein considered Lower Cambrian because of its structural and stratigraphic relationships to the overlying fossiliferous Antietam quartzite.

The Harpers phyllite overlies the Weverton quartzite in normal sequence along the eastern slopes of Catoctin Mountain. The formation extends from the northern end of the area southward to 3 miles north of Leesburg,

Virginia, where it is cut off by the Triassic border fault. With the exception of occasional road cuts, exposures of the Harpers are very poor so that most of the formation has been mapped on the basis of chips of phyllite scattered in fields. Much of the Harpers is covered by alluvial mountain wash in the northern portion of the area.

The Antietam quartzite overlies the Harpers in normal sequence throughout much of the area. However, between Thurmont and Mount St. Marys in northern Maryland, the Harpers is overlain by the Frederick limestone. In contrast, just north of the Potomac River the Harpers is overlain by the Tomstown dolomite. These anomalous stratigraphic relationships will be discussed later.

The stratigraphy of the Harpers needs revision. As previously stated, what has been mapped as Loudoun on the eastern side of Catoctin Mountain (Stose and Jonas, 1938) is in reality a portion of the lower Harpers and upper-most Weverton.

The Harpers is a finely laminated, argillaceous phyllite, light bluish-gray (5B7/1) on fresh surfaces and dusky-yellow (5Y6/4) when weathered. The phyllite is characterized by light-gray (N7) and dusky blue-green (5BG3/2) laminae. The basal portion of the Harpers is characterized by a distinctive pale-purple (5P6/2) phyllite occurring in irregular lenses. The geographic position of these lenses is shown in Figure 8.

In thin section the Harpers is a strongly schistose rock composed of quartz, chlorite, and sericite with minor amounts of albite (Ab98-96) and magnetite. The megascopic banding which denotes bedding is composed chiefly of quartz in the light-gray bands and of chlorite and sericite in the dusky blue-green bands. Flow cleavage is made very distinct by elongate aggregates of quartz grains and a parallel alignment of the chlorite and sericite. Chlorite occurs in the groundmass as finely divided grains parallel to the trace of the flow cleavage, and as porphyroblasts, many of which are transverse to the trace of the flow cleavage of the phyllite (Pl. 2, fig. 2).

The pale-purple phyllite, occurring in lenses at the base of the Harpers is identical in thin section with the phyllite at the base of the Loudoun. Both are considered metamorphosed rhyolite tuffs.

No complete section of the Harpers is exposed in the area mapped by the writer, and the obliteration of bedding by flow cleavage makes measurement of sections impossible. Cloos (1951, p. 33) estimated 3100 feet of Harpers near Waynesboro, Pennsylvania, where the phyllite occurs in the crestal position of the South Mountain fold so that flow cleavage makes a large angle with bedding, and distortion of thickness is considerable.

Antietam Quartzite

The Antietam quartzite was first differentiated but not named by Keith (1892, p. 365). It was later named in separate publications by Keith (1893, p. 335) and by Williams and

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PLATE 2.—PHOTOMICROGRAPHS OF LOUDOUN PHYLLITE, HARPERS PHYLLITE, TOMSTOWN DOLOMITE, AND TRIASSIC DIABASE

FIGURE 1.—CROSSED NICOLS

Fracture cleavage in the Loudoun phyllite showing crenulation of flow cleavage and slippage along fracture-cleavage planes.

FIGURE 2.—PARALLEL LIGHT

Harpers phyllite showing porphyroblastic chlorite in matrix of sericite, quartz, and magnetite. Mineral cleavage of chlorite nearly perpendicular to that of flow cleavage of the rock.

FIGURE 3.—CROSSED NICOLS

Tomstown dolomite showing porphyroblasts of dolomite in matrix of granoblastic dolomite.

FIGURE 4.—CROSSED NICOLS

Ophitic texture of Triassic diabase dike showing labradorite laths with interstitial augite. (Photomicrographs by Miner B. Long).

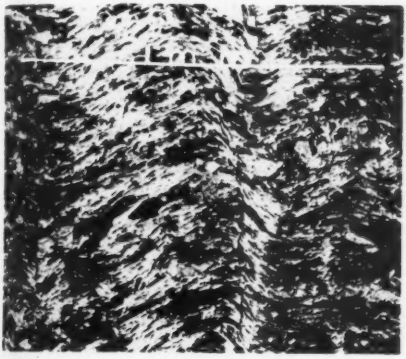

FIGURE 1

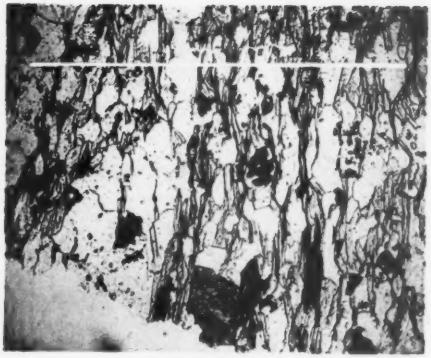

FIGURE 2

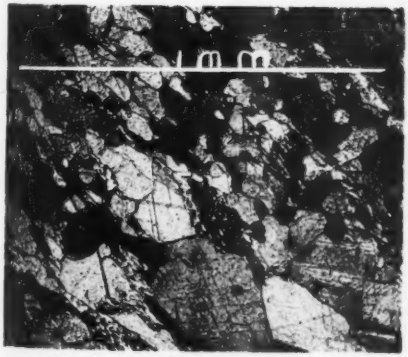

FIGURE 3

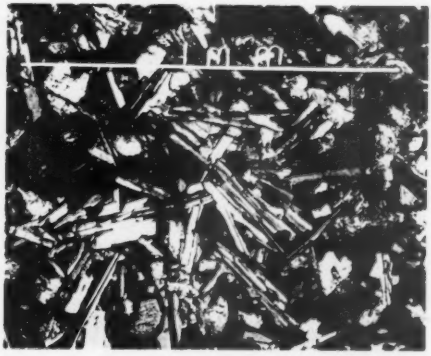

FIGURE 4

PHOTOMICROGRAPHS

FIGURE 2

FIGURE 1

FOLDS IN WEVERTON QUARTZITE

Clark (1893, p. 68). In 1894 Keith (p. 3) further described the formation stating that it was best shown on the tributaries of Antietam Creek in Washington County, Maryland.

Although no fossils were found by the writer, the Antietam has yielded a few fossil fragments from localities in Maryland and southern Pennsylvania, which indicates Lower Cambrian. In 1896 Walcott (p. 25) recorded fragmentary remains of *Olenellus* sp., *Camarella minor* (*Obolella minor* Walcott), *Hyolithes communis* Billings, and the supposed worm boring *Scolithus linearis* Haldemann (Bassler, 1919, p. 56; Resser, 1938, p. 5; Amsden, 1951, p. 102). In the area mapped by the writer, fragmentary fossils have been found in the Antietam 2 miles east of Jefferson, Maryland (Stose and Stose, 1946, p. 42).

The Antietam quartzite is poorly exposed along the eastern foothills of Catoctin Mountain in a narrow belt of minor ridges about a quarter of a mile wide. It crops out from just south of Church Valley in Loudoun County, Virginia, northward to Tuscarora Creek northwest of Frederick, Maryland. For a distance of 1½ miles northward from the Potomac River the Harpers is mapped in contact with the Tomstown, and the Antietam is absent; the implication of this anomalous stratigraphic relationship will be discussed later.

The Antietam is a uniformly fine-grained quartzite with thin argillaceous partings. Fresh surfaces are greenish-gray (5GY 6/1), but on exposure it becomes light olive-gray (5Y 6/1). Throughout most of the area, flow cleavage has obliterated bedding. At a few localities bedding can be seen as a faint argillaceous parting, but it is difficult to distinguish from sericitic growth along flow-cleavage planes where the angle between bedding and flow cleavage is small.

In thin section the Antietam is composed of uniformly fine, anhedral, interlocking grains of quartz with lesser amounts of albite and microcline. Minor magnetite and chlorite are also present. Cleavage dominates the rock and is expressed as an elongation of the quartz grains as well as an alignment of sericite and minute, crushed and recrystallized quartz grains.

No complete section of the Antietam is exposed in the area mapped by the writer but the thickness, estimated from scattered exposures in Frederick County, is 300 feet (Stose and Stose, 1946, p. 42).

Tomstown Dolomite

The Tomstown formation was named by Stose (1906, p. 208) for exposures at the village of Tomstown near Chambersburg, Pennsylvania, where a dolomite sequence overlies the Antietam quartzite. No fossils were found by the writer in the area mapped, but in Maryland the formation carries fragmentary remains of olenellid trilobites, the brachiopod *Kurtogina*, and the mollusk *Salterella*, indicating Lower Cambrian (Walcott, 1896, p. 25; Bassler, 1919, p. 66; Resser, 1938, p. 6–7; Amsden, 1951, p. 102).

The Tomstown formation extends along the eastern foothills of Catoctin Mountain from west of Frederick, Maryland, to just north of Church Valley in Loudoun County, Virginia. The dolomite lies between the ridges of Antietam quartzite and the Triassic border fault. The formation is poorly exposed, and most of the dolomite has been mapped on the basis of its residual clays.

Most of the Tomstown is cut off by the Triassic border fault so that only the lowest portion of the formation is exposed in gradational contact with the underlying Antietam quartzite. The lowermost beds are composed of a light bluish-gray (5B 7/1) knotty dolomite interbedded with paper-thin argillaceous partings. The rock weathers to a moderate reddish-brown (10R 4/6) residual clay which in places is

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PLATE 3.—FOLDS IN WEVERTON QUARTZITE

FIGURE 1.—TIGHTLY FOLDED ANTICLINE

Cleavage transects bedding at high angle near crest of fold and nearly parallels bedding in upper limb; beds thicken in crest, thin out in limbs.

FIGURE 2.—CREST OF ANTICLINE OF LEDGE-MAKER MEMBER OF WEVERTON

covered by wash from the adjacent mountain slope.

In thin section the Tomstown is composed of euhedral porphyroblasts of dolomite in a groundmass of recrystallized, interlocking grains of anhedral dolomite with minor amounts of calcite (Pl. 2, fig. 3). The dolomite porphyroblasts contain abundant inclusions of anhedral quartz, indicating large-scale replacement. The thin argillaceous partings are composed of sericite with small pods of interlocking grains of quartz and minor amounts of microcline.

Frederick Limestone

The Frederick limestone was named by Bassler (1919, p. 115) for the platy-weathering "building or fence stone" common around Frederick, Maryland. Bassler (1919, p. 117; *see also* Keyes, 1890, p. 32) designated the Frederick limestone as Chazyan or early Mohawkian; however, later work has indicated it to be Upper Cambrian (Stose and Stose, 1946, p. 46). Ulrich and Cooper (1938, p. 44, 76–77) have identified the brachiopods *Xenorthis stosei* (Bassler), *X. jonasae* Ulrich and Cooper, and *X. sequicostellata* Ulrich and Cooper. In addition, Wilson (1952, p. 137) has identified the trilobites *Plethometopus* sp., *Stenopilus* sp., *Euptychaspis* sp., and *Acidaspis ulrichi* Bassler. Both Cooper (*in* Stose and Stose, 1946, p. 46) and Wilson (1952, p. 317) assign an Upper Cambrian age to the Frederick limestone.

The formation is exposed in three areas along the eastern side of Catoctin Mountain. Between Thurmont and Mount St. Marys, Maryland, the Frederick limestone overlies the Harpers phyllite. In contrast, along U. S. Highway 40 west of Frederick and in a restricted area immediately south of the Potomac River, the Frederick limestone overlies the Tomstown dolomite. These anomalous stratigraphic relationships will be discussed later.

The Frederick limestone is a distinctive, thin-bedded, dark-gray (N3) limestone in layers half an inch to 1½ inches thick, separated by thinner light-gray (N7) silty partings an eighth of an inch to half an inch thick. On weathering it breaks into slabby, medium light-gray (N6) layers with irregular bedding surfaces. The limestone is characterized by numerous intersecting minute fractures filled with secondary calcite.

In thin section, the limestone layers are composed of calcite and dolomite porphyroblasts in a groundmass of granular, interlocking, recrystallized grains of anhedral calcite. The light-gray silty partings are predominately carbonate and quartz with occasional fragments of tourmaline.

No continuous section of the Frederick limestone is exposed in the area mapped by the writer, but an estimated thickness of 480 feet is based on scattered exposures in the Frederick Valley (Stose and Stose, 1946, p. 43–44).

Stratigraphic Relationship of Upper Chilhowee Group and Overlying Cambrian Carbonate Rocks

The known stratigraphic sequence of the upper Chilhowee group and the overlying carbonate rocks in the South Mountain and Great Valley area differs somewhat from those in the vicinity of Catoctin Mountain and the Frederick Valley. The sequences are:

South Mountain Great Valley Area		*Catoctin Mountain Frederick Valley Area*	
U€	Conococheague limestone	U€	Frederick limestone
M€	Elbrook limestone		
L€	Waynesboro dolomite and shale		
	Tomstown dolomite	L€	Tomstown dolomite (locally absent)
	Antietam quartzite		Antietam quartzite (locally absent)
	Harpers phyllite		Harpers phyllite

The stratigraphic sequence from the Harpers phyllite to the overlying carbonate rocks is not everywhere the same along the eastern slope of Catoctin Mountain. In the northern portion of the area, Harpers is overlain by the Frederick limestone. Farther south in the foothills west of Frederick, Maryland, the Harpers is overlain by the Antietam. Just north of the Potomac River, the Harpers is overlain by the Tomstown, whereas south from the Potomac River in Loudoun County, Virginia, the Harpers is overlain by the Antietam quartzite.

Any one of several hypotheses may explain these anomalous stratigraphic relationships: (1) faulting; (2) facies change, or (3) one or more unconformities.

If structural control is eliminated, these anomalous stratigraphic relationships may be explained by facies changes or one or more unconformities within the section; however, the

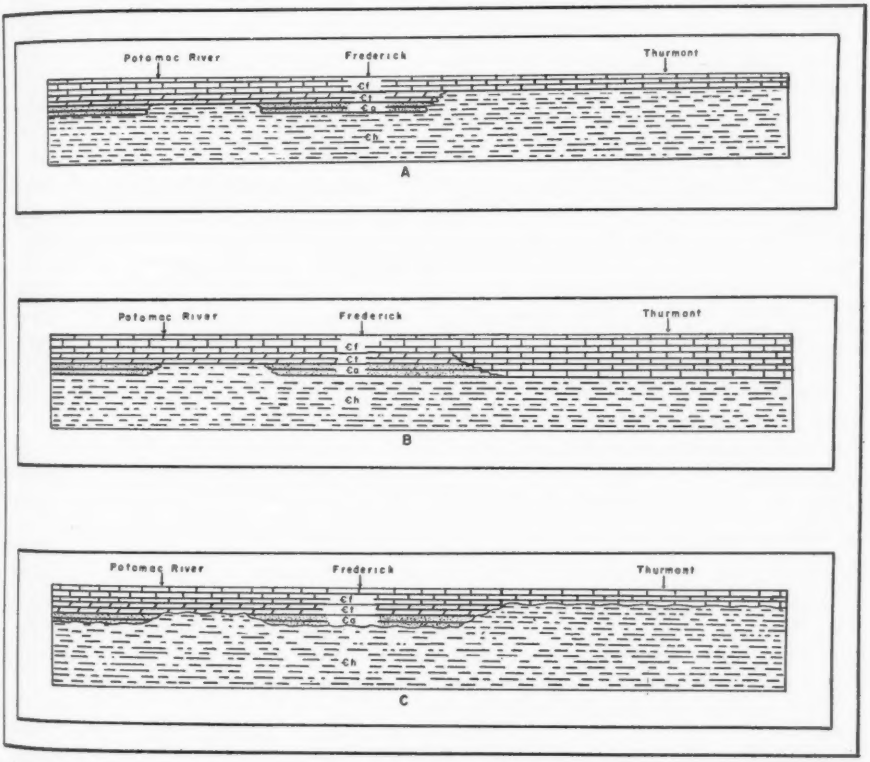

FIGURE 3.—HYPOTHETICAL STRATIGRAPHIC RELATIONSHIPS OF UPPER CHILHOWEE GROUP TO OVERLYING CAMBRIAN CARBONATE ROCKS

Harpers phyllite *eh*, Antietam quartzite *ea*, Tomstown dolomite *et*, Frederick limestone *ef*

No faults have been observed between any of the upper Chilhowee group and the overlying Cambrian carbonate formations. Between Thurmont and Mount St. Marys, Maryland, the Frederick limestone overlies the Harpers phyllite in a tight, gently plunging syncline. That the Frederick and Harpers are in fault contact in this vicinity seems unlikely since this would date the faulting as pre-folding, a condition unknown in the South Mountain province. Similarly, the anomalous stratigraphic relationships west of Frederick, Maryland, and in the vicinity of the Potomac River show no evidence of fault contacts.

writer can say with no assurance which of these factors is the controlling one.

In the vicinity of the Potomac River the sandy Antietam lithology may give way to the Harpers phyllite, whereas northward from Frederick, Maryland, the Antietam sands and Tomstown dolomite give way to the Harpers phyllite lithology (Fig. 3A). It is possible that the Frederick limestone may thicken at the expense of the Harpers phyllite in the northern portion of the area (Fig. 3B). In either hypothesis the ages cannot be ascertained since the formations have been mapped solely on the basis of lithology without the aid of fossils.

Figure 3C assumes that a post-Harpers pre-Antietam unconformity exists, placing the Tomstown over the Harpers in the vicinity of the Potomac River, and the Frederick over the Harpers in the northern portion of the area. This interpretation suggests that following the deposition of the Harpers the area was uplifted and eroded to produce an irregular surface. The Antietam and Tomstown were then deposited in lower areas as the sea transgressed. A variation of this hypothesis is that two unconformities are present; one between the upper Cambrian Frederick limestone and the underlying Cambrian formations, and a second unconformity within the lower Cambrian as just described.

Grove Limestone

The Grove limestone was named by Stose and Jonas (1935, p. 565) for the thick-bedded, pure limestone exposed at Grove Station southeast of Frederick, Maryland. According to Stose and Stose (1946, p. 50) Ulrich and Foerste have identified the cephalopods: *Ectenolites* sp., *Levisoceras* sp., *Walcottoceras* sp., *Cameroceras* sp., *Clarkoceras* sp.

Ulrich and Cooper (1938, p. 32–33, 36, 38, 42, 87, 89, 203, 212, and 220) have identified the brachiopods: *Clarkella depressa*, *Acrotreta* sp., *Apheoorthis* sp., *Glyptotrophia marylandica*, *Nanorthis difficiles*, *Syntrophina impressa*.

Bridge identified a few gastropods as *Sinnopea* sp. (Stose and Stose, 1946, p. 51).

According to Ulrich and Foerste, the fauna indicates that the Grove limestone is a correlative of the Chepultepec of the southern Appalachians and the Gasconade of Missouri, both lowermost Ordovician (Stose and Stose, 1946, p. 50). Ulrich and Cooper (1938) also assign a lowermost Ordovician age (Upper Ozarkian of Ulrich) to this fauna.

Throughout much of the Frederick Valley the Grove limestone is mapped in tight synclines overlying the Frederick limestone (Stose and Jonas, 1938). In the area mapped by the writer, a thick-bedded limestone with massive beds of dolomite lies in fault contact with the Harpers phyllite northwest of Lewistown, Maryland. Both the writer and Stose and Jonas (1938) map this as Grove limestone.

The Grove limestone is a thick-bedded, pure limestone with massive beds of fine-grained dolomite in the lower part. The base of the formation is characterized by thin layers of limestone with scattered grains of quartz. In the LeGore quarry near Frederick, Maryland, 590 feet is exposed (Stose and Stose, 1946, p. 47).

Triassic Diabase

A think sequence of carbonate conglomerate, red arkosic sandstone and shale of the Newark series associated with diabase sills and dikes lie to the east of the Triassic border fault outside the area mapped by the writer. On the eastern slope of Catoctin Mountain, northwest of Feagaville, Maryland, a Triassic diabase dike is traceable for 1½ miles. It strikes N.5°E., thus transecting the regional trend at a small angle.

The diabase is fine-grained, dark gray (N3) on fresh surfaces, and weathers to a light olive gray (5Y 6/1). In thin section, it shows an ophitic texture of laths of labradorite (Ab 52–47) with interstitial augite (Pl. 2, fig. 4). Magnetite is the only accessory mineral, and augite is partially altered to chlorite. A chemical analysis of the diabase from Frederick County is given by Stose and Stose (1946, p. 97).

Quaternary Alluvium

Alluvial mountain wash at the mouths of gorges on the east slope of Catoctin Mountain and the floor of the Frederick Valley is composed of well-rounded boulders of Weverton quartzite and Loudoun conglomerate. The largest alluvial deposit is at the mouth of Hunting Creek, and spreads eastward from the foot of Catoctin Mountain for about 3 miles. Other alluvial deposits occur at the mouths of Little Hunting, Fishing, and Little Tuscarora creeks, and the north branch of Owens Creek. These alluvial deposits have been entrenched by the streams that formed them. The thickness of the alluvial wash is estimated to be as great as 250–300 feet. Wells drilled at Thurmont and Blue Mountain reached 200 and 150 feet respectively without passing through the mountain wash.

Alluvium is also present on the flood plains of all the larger streams in the area but has been mapped only along the Potomac River because of the restricted extent of the deposits along the other streams.

Metamorphism

The Blue Ridge province of Maryland and northern Virginia may be classified according to mineral facies, as defined by Eskola (1920, p. 146), or referred to a metamorphic zone after the manner first utilized by Barrow (1893). The rocks investigated at Catoctin Mountain belong to the greenschist facies (Eskola, 1939) and, since they lack biotite, belong to the muscovite-chlorite subfacies suggested by Turner (1948, p. 96).

The Harpers phyllite is typical of the argillaceous rocks metamorphosed chiefly through the action of shearing stress (dynamic metamorphism) (Harker, 1939, p. 209–214). Recrystallized detrital quartz occurs as aggregates of elongated grains parallel to the trace of flow cleavage. The argillaceous constituents, converted to sericite and chlorite, are aligned parallel to the flow cleavage. Occasionally, chlorite occurs as porphyroblasts transverse to the trace of flow cleavage (Pl. 2, fig. 2). Scotford (1951, p. 58) has shown that a uniform albite (Ab_{98-96}) occurs in all the samples taken from the Harpers phyllite from Sugarloaf Mountain, east of Catoctin Mountain, to as far west as South Mountain. The presence of albite in the Harpers phyllite at Catoctin Mountain confirms this. Whether the albite is detrital or the product of metamorphism is undetermined. The writer prefers the latter interpretation because: (1) there is abundant evidence that albite, in contrast to potash and calcic feldspars, is stable under varying conditions of stress and temperature and should be considered as typical a stress mineral as the associated muscovite and chlorite (Harper, 1939, p. 212); and (2) it seems improbable that a uniform albite would supply detritus over such a large area.

The metabasalt flows of the Catoctin formation also have been metamorphosed under conditions of low heat and pressure and belong to the greenschist facies. Five Rosiwal analyses in addition to a chemical analysis indicate that the rock has the composition of a normal basalt. Epidote and chlorite seem to have been derived from the original pyroxene and more calcic plagioclase.

The quartzites and carbonate rocks show no evidence of a change in mineralogy due to metamorphism; only recrystallization and shearing of the original minerals are evident.

Megascopic Structure

Introduction

The writer has recorded all megascopic structures in order to describe the deformation pattern more accurately. The geographic orientation of structure elements are read from a 360° scale. Some structural features are presented in statistical diagrams (Figs. 4*a–f*, 6*a*, *b*). These are the lower half of an equal-area net projection oriented into geographic position.

Primary Structures

BEDDING: Bedding has been superseded by flow cleavage as the dominant structure in most formations. In the Harpers phyllite and basal phyllite member of the Loudoun, as well as in the fine-grained, argillaceous quartzite of the Antietam, flow cleavage is so prominent that bedding has been almost entirely obliterated. In contrast, in the Weverton quartzite, bedding remains the dominant structure, and flow cleavage can be seen only on close inspection.

In the Tomstown and Frederick limestones and dolomites, bedding becomes highly crenulated, but it is usually recognizable by numerous shaly partings separating the carbonate layers.

In the Catoctin formation, flow cleavage is very prominent. The only evidence of primary structure the author found was occasional amygdaloidal layers indicating the tops of flows.

Secondary Structures

FLOW CLEAVAGE: In the incompetent phyllites and fine-grained, argillaceous quartzites flow cleavage is easily distinguished by the pronounced parallel alignment of the platy minerals.

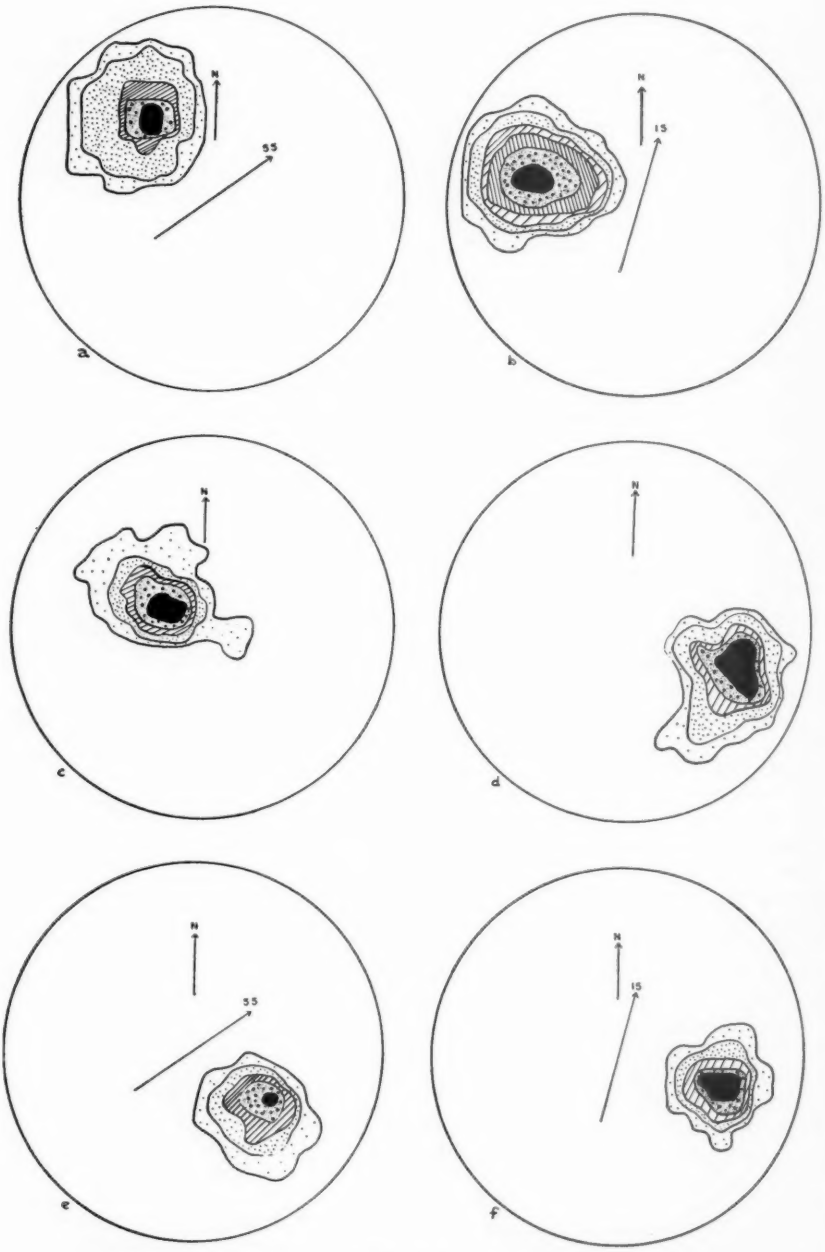

FIGURE 4

In the more competent Weverton quartzite, flow cleavage is visible as an orientation of elongated quartz grains and parallel alignment of platy minerals. Where the crests of folds are visible, cleavage transects bedding at high angles and often almost obliterates the primary structure.

In the Frederick and Tomstown limestones and dolomites, cleavage is visible in the shaly partings due to parallel alignment of the platy minerals. In the carbonate layers the cleavage is visible as elongated oöids.

Flow cleavage is the dominant structure of the South Mountain fold. It forms an axial-plane cleavage which dips uniformly eastward and fans gently westward and upward. The constant eastward dip of the flow cleavage is shown (Fig. 4*a*, *b*). Since the trend of Catoctin Mountain changes from N.15°E. in the southern portion to N.55°E. north of Owens Creek near Thurmont, Maryland, separate plots have been made to show the dependence of the strike of the flow cleavage on the general trend of the area.

Near the axial plane of a fold, flow cleavage parallels the axial plane, but it fans slightly in the limbs. Thus, in the west limb of an overturned anticline the bedding dips more steeply than the cleavage, while in the east limb cleavage dips more steeply than bedding (Nevin, 1949, p. 171).

In the Catoctin Mountain area, flow cleavage dips uniformly eastward and fans gently westward in the small, tight anticlines seen in the field. This relation is useful for locating bedding. In general, where cleavage dips steeply eastward (>45°) the bedding is at a lower angle, and the beds are right-side up; where cleavage dips less steeply eastward the bedding is nearly vertical or overturned.

Not only does the flow cleavage fan in the tight folds observed in the field, but it also fans in the entire South Mountain anticlinorium. This is substantiated by the contrast in position of the poles of flow cleavage plotted at Catoctin Mountain in the eastern and upper limb with attitudes measured by Cloos (1941; 1947; 1951) in the western over-turned limb at South Mountain (Fig. 4*a–c*). Whereas flow cleavage at Catoctin Mountain dips consistently eastward and averages about 50°, that at South Mountain dips less steeply eastward and averages about 20°. This relationship is portrayed in the generalized cross section of the South Mountain fold (Fig. 2).

FRACTURE CLEAVAGE: Fracture cleavage transects bedding and flow cleavage in all formations. It is most prominent in the Harpers phyllite and the basal phyllite member of the Loudoun and is expressed as a closely spaced fracturing or as a crenulation of the flow cleavage so that fracture cleavage often forms the axial plane of minute chevron folds. At some localities the flow-cleavage planes are displaced by slip on fracture cleavage surfaces.

Although only 53 fracture cleavage readings were taken, a statistical plot of the poles of these planes (Fig. 4*d*) shows a constant west dip varying from 25° to 75°. The strike of the intersection of fracture and flow cleavage does not consistently represent the direction of fold axes. At places this intersection is parallel to the strike of small folds but more often the intersection deviates in strike about 10° from the fold axes.

LINEATION: In describing lineation Sander's (1930, p. 119) co-ordinate system is used: *b* is the fold axis, *a* is perpendicular to *b* in the movement plane; and *c* is perpendicular to *ab*. The *ab* plane represents the flow-cleavage plane.

FIGURE 4.—EQUAL-AREA PROJECTIONS

a. Poles of 314 flow-cleavage planes plotted from north of Owens Creek. General trend N.55°E.; average dip of flow cleavage 50°E., maxima 3, 7, 13, 15, and 17 per cent.

b. Poles of 506 flow-cleavage planes plotted from south of Owens Creek. General trend N.15°E.; average dip of flow cleavage 50°E.; maxima 3, 7, 10, 13, 15, and 17 per cent.

c. Poles of 100 flow-cleavage planes plotted from South Mountain. Average dip of flow cleavage 20°E., maxima 4, 7, 12, 20, and 28 per cent (Cloos, 1941, 1946, 1947, 1951).

d. Poles of 53 fracture-cleavage planes. Maxima 1, 4, 7, 12, and 14 per cent.

e. Projection of 97 *a* lineations plotted from north of Owens Creek. General trend N.55°E.; maxima 3, 8, 12, 18, and 24 per cent.

f. Projection of 204 *a* lineations plotted from south of Owens Creek. General trend N.15°E.; maxima 3, 8, 12, 18, and 24 per cent.

As defined here, lineation includes all linear structures without regard to origin (Cloos, 1947, p. 1).

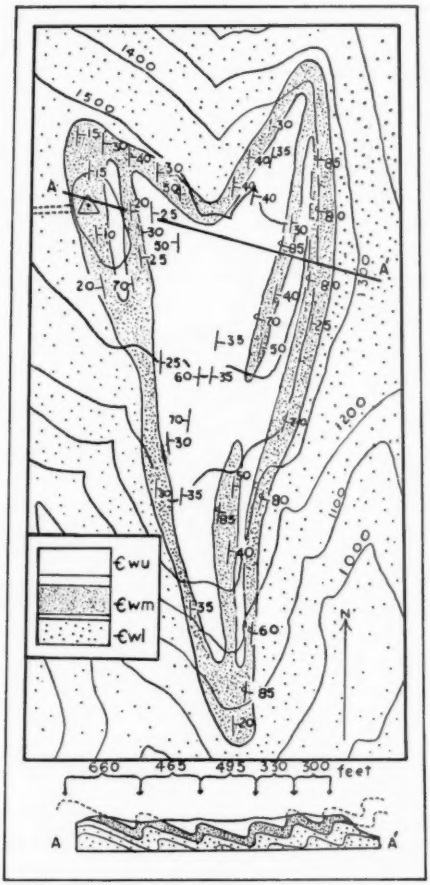

FIGURE 5.—GEOLOGIC MAP OF AREA IMMEDIATELY EAST OF HAMBURG FIRE TOWER

Profile indicates distance between crests of anticlines.

At Catoctin Mountain the following kinds of lineations are found: (1) axes of small folds or crenulations, (2) intersection of bedding and flow cleavage, (3) intersection of flow and fracture cleavage, and (4) lineation in the *a* direction due both to crenulation of the flow-cleavage plane and stretching of minerals in the *a* direction.

Small folds or crenulations are prominent in the Harpers phyllite and basal phyllite member of the Loudoun and are convenient for measuring fold axes.

Similarly, the intersection of flow cleavage and bedding parallels the fold axes. Along this intersection secondary quartz may be very prominent.

Fold axes throughout the area are essentially horizontal. Plunges of more than 10° are uncommon in the competent ledge-maker member of the Weverton. In the incompetent Harpers phyllite and basal phyllite member of the Loudoun, fold axes are more varied and plunge locally as much as 40°.

The intersection of fracture and flow cleavage forms a lineation due to minute crenulations of the flow-cleavage plane. As stated above, this intersection is not a dependable indicator of the direction of fold axes.

Lineation in the *a* direction is the dominant lineation of the area. It is present in all the rocks of the South Mountain fold, from the Precambrian granodiorite and biotite granite gneiss core as far west as the lower Ordovician limestones (Cloos, 1947, p. 845). At Catoctin Mountain it consists of an elongation of minerals down the dip of the flow-cleavage plane or a crenulation of the flow-cleavage plane in *a*. Figures 4*e* and *f* show the remarkable constancy of this structure throughout the entire area.

FOLDS: The folds in the area are asymmetrically overturned to the west. They are shear folds where the principal movement has been along flow-cleavage planes rather than bedding planes. Thinned limbs with thickened crests and troughs are the result of this movement. Detailed mapping in a well-exposed area gives a rather accurate representation of the distances between the crests of anticlines. Figure 5 is a detailed map of the area immediately east of the Hamburg fire tower where both the ledge-maker member and the overlying ferruginous quartz conglomerate of the Weverton are exposed. Erosion has cut downward on the borders of this plateaulike area so that the trace of the ledge-maker member can be mapped in the valley wall as well as the beds immediately below the ledge-maker member. Section A-A′ (Fig. 5) shows the approximate position

of the crests of six anticlines, and the cross section is an accurate representation of the folding within a limited area. The distance between crests of anticlines ranges between 660 and 300 feet. Since there has been movement along flow-cleavage planes resulting in flowage toward the crests of anticlines and the troughs of synclines no estimate of the amount of lateral shortening involved can be given. The wave length of folding determined in this limited area is not constant for the entire Catoctin Mountain region. In many areas the trace of the ledge-maker member can be mapped along the valley walls for several hundred yards, and the beds dip uniformly eastward with no indication of folding. Similarly, road cuts along U. S. Highway 40*A* and 340 expose a uniformly eastward-dipping sequence.

JOINTS: The writer measured orientations of about 900 joints and plotted their poles in an equal-area net. No recognizable pattern results if all the joints are plotted in one diagram. If one considers, however, the competency of the formations involved, the frequency of exposures, and most important, the position of the joints with respect to the general trend of the area, several strong maxima appear. Since the phyllitic, volcanic, and carbonate rocks are poorly exposed units of varying competency, their joint readings were disregarded. The ledge-maker member of the Weverton forms a well-exposed unit of uniform competency suitable for the analysis of joints; therefore, only joints from the ledge-maker member are considered.

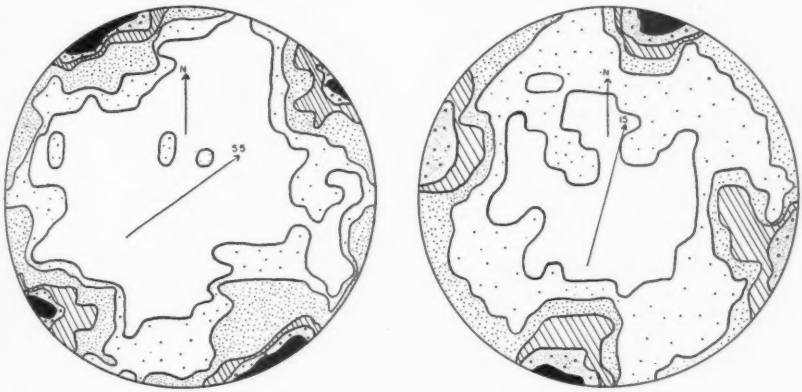

FIGURE 6.—EQUAL-AREA PROJECTIONS

a. Projection of 262 joints from ledge-maker member of Weverton plotted from north of Owens Creek. General trend N.55°E.; maxima 1, 2, 5, 10, and 14 per cent.

b. Projection of 563 joints from ledge-maker member of Weverton plotted from south of Owens Creek. General trend N.15°E.; maxima 1, 3, 7, 10, and 14 per cent.

Separate plots (Fig. 6*a*, *b*) show the relation of the joint pattern to the two general trends of Catoctin Mountain. The two plots reveal the following maxima: cross joints (*ac* joints), strike joints (*bc* joints), and two sets of conjugate joints, one of which intersects in *b* (*hol* joints), the other in *a* (*okl* joints).

Cross joints form 14 and 13 per cent maxima in Figures 6*a* and *b* respectively. They are extremely common, and many are accompanied by slickensides and feather joints zones which indicate minor displacements perpendicular to fold axes.

Strike joints form a 14 and 10 per cent maxima in Figures 6*a* and *b* respectively. These joints also show slickensiding, but no systematic investigation of the direction of movement was undertaken.

Of particular interest are the conjugate shear joints which intersect in *a* or *b* (*okl* and *hol* joints respectively). The strike of these joints parallels that of the *ac* or *bc* joints, and their dip averages about 75°. The meaning of these two

sets of conjugate joint sets is not clear, but the writer calls attention to them in the hope that further studies will be undertaken. As a general rule *okl* and *hol* joints are associated with tightly folded beds, whereas vertical *ac* and *bc* joints are most commonly associated with flat or gently dipping beds.

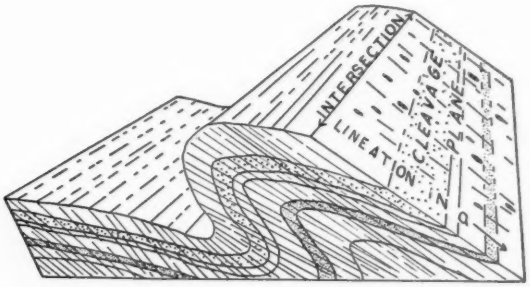

FIGURE 7.—DEFORMATION PLAN

It seems unlikely that large, open, quartz-filled joints could survive in the plastic state of deformation. Since the South Mountain fold seems to show the imprint of plastic deformation (Cloos, 1947, p. 913), joints either must have originated in the late stage of deformation or must be attributed to post-deformation rupture. The writer feels that the first alternative is correct because of: (1) the striking tendency for the joint maxima (Fig. 6*a*, *b*) to be oriented with respect to the two general trends of the area, and (2) the majority of joints in any exposure which can be related to the deformation axes *a*, *b*, and *c*.

For these reasons the author is inclined to believe that jointing, while obviously the product of the late stages of deformation, is not a completely post-deformational feature; if so, the joints would not be so closely related to the deformation plan.

Deformation Plan

Catoctin Mountain is a portion of a larger structural province termed the South Mountain fold. From the Triassic border fault on the east, to the Ordovician limestones of the Great Valley on the west, the same deformation plan exists. Figure 7 is a schematic presentation of the typical deformation plan of the South Mountain area: folds are asymmetrically overturned to the west; flow cleavage dips eastward and fans gently toward the west; the principal movement has been along flow-cleavage planes so that shear folds result with thickened crests and troughs and thinned limbs; fold axes are subhorizontal; lineation in *a* results from a crenulation of the flow cleavage and a stretching of minerals down the dip of the flow-cleavage plane; and the intersection of flow cleavage and bedding produces a lineation in *b*.

STRUCTURAL INTERPRETATION

Introduction

Keith (1893) and Stose and Jonas (1938) mapped Catoctin Mountain as a lightly folded syncline overturned to the west. By this interpretation the Loudoun formation is exposed on both the east and west sides of the mountain and encloses the overlying Weverton. To the east the Loudoun is mapped in fault contact with the Harpers. Stose and Jonas (1938) mapped what is here termed the ledge-maker member of the Weverton as tightly folded synclines. As a result their map pattern shows a series of unconnected lens-shaped bodies representing tightly folded synclines of the ledge-maker member.

The writer concludes that Catoctin Mountain forms a tightly folded normal sequence which dips eastward and thus forms the upper limb of the South Mountain fold. This interpretation necessitates a revision of the stratigraphy. What has been mapped by Stose and Jonas (1938) as Loudoun on the east side of

Catoctin Mountain is actually a zone including the uppermost Weverton and the basal portion of the Harpers.

Map Pattern

The interpretation of Catoctin Mountain as a tightly folded, eastward-dipping normal sequence is best brought out by the outcrop pattern of the ledge-maker member of the Weverton mapped in the northern portion of the area. The key to the area lies in mapping the trace of the ledge-maker member along the valley walls. In all the valleys where the ledge-maker member is present the trace of these beds can be followed as they dip eastward under the Harpers in normal sequence.

Along the east slope of the mountain the folds can rarely be mapped because they are obscured by float. However, in the high, flat, plateau-like areas erosion has exposed the crests of many anticlines. The trace of these folded beds can also be mapped along the valley wall. Thus, the ledge-maker member is similar to a sheet of corrugated metal which is only partially exposed so that the crests of folds can only occasionally be mapped but the trace of the folded bed is always present in the valley wall.

Structure of Catoctin Mountain

Just southwest of Emmitsburg, Maryland, Catoctin Mountain is cut off by the Triassic border fault; southward the mountain trends N.55°E. The valley immediately east of the mountain is a gently plunging syncline enclosing the Harpers and overlying Frederick limestone. Fold axes in the Harpers plunge as steeply as 40°N.E. just south of Owens Creek but become subhorizontal a few miles farther north. The contact between the Frederick and the Harpers can be seen in Owens Creek 200 yards east of U. S. Highway 15. Bedding is nearly vertical in the Frederick limestone and cleavage dips about 20°E., indicating that the exposure is in the lower limb of an overturned anticline.

The best exposures at Catoctin Mountain are along State Highway 81 and the Western Maryland Railroad in the gorge of Owens Creek. In a diagram of the structure through Owens Creek gorge, Stose and Stose (1946, p. 35, Fig. 9), show the Weverton in an overturned syncline in contact with the Loudoun to the east. The writer found instead a normal eastward-dipping sequence where the trace of each mappable unit follows along the valley wall and dips eastward under the next succeeding younger unit. Cleavage dips more steeply than bedding throughout the entire gorge. Cross bedding and graded bedding also verify that the sequence is normal. What has been mapped as Loudoun at the east entrance to the gorge is actually the lower portion of the Harpers.

The ledge-maker member is present from the northern end of the mountain to just north of Tuscarora Creek near Yellow Springs, Maryland. Where Hunting, Fishing, Little Fishing, and Owens creeks have cut through the mountain, the eastward-dipping trace of the ridge maker can be seen in the valley wall. In the high, plateau areas just east of Hamburg Fire Tower, at Bob's Hill, and in the vicinity of Dry Branch, the tightly folded crests of anticlines of the ledge-maker member can be mapped. Small isolated bodies farther west (*i.e.*, Cat, Salamander, and Black rocks) were mapped mainly as loose blocks which must be interpreted as isolated infolds of the ledge-maker member since no connecting trace in the valley walls was found between these isolated bodies and the ledge-maker member to the east.

South of Tuscarora Creek deep gorges expose the crests of anticlines of both the conglomerate and underlying basal phyllite members of the Loudoun. Along the road leading northward from Gambrill State Park and along the steep cliffs west of this road, the Loudoun conglomerate is well exposed. Throughout this entire area cleavage dips more steeply than bedding indicating that the beds are on the upper limb of the South Mountain anticlinorium.

From U. S. Highway 40 to just north of the Potomac River the mountain forms a simple, uniformly eastward-dipping normal sequence. The road cut on the U. S. Highway 340 west of Jefferson shows this relationship very clearly. On the Frederick County map Stose and Jonas (1938) show the Loudoun in an overturned syncline in fault contact with the Harpers to the

east. The east entrance to the road cut exposes pale-purple phyllite in gradational contact with an 87-foot section of the finely laminated Weverton quartzite. Just west of the road cut, a few yards north of the highway, a ledge of Loudoun conglomerate is exposed. Throughout the entire road cut, cleavage dips more steeply than bedding. In addition, the phyllites at the east entrance to the road cut are in gradational contact with the finely laminated quartzites of the Weverton. Therefore, what has been mapped as overturned Loudoun at the east entrance of the road cut is actually the tuffaceous phyllite at the base of the Harpers. Thus, the road cut shows a normal eastward-dipping sequence of lower Harpers, Weverton, and a partial section of the Loudoun conglomerate.

In the vicinity of the Potomac River a normal fault is present. The fault begins just north of State Highway 464 where the basal phyllite member of the Loudoun is in contact with the ferruginous quartz conglomerate beds of the upper Weverton. The displacement on the fault reaches a maximum at the Potomac River where the Catoctin formation is against the Harpers phyllite. Displacement gradually decreases southward so that a normal sequence of eastward-dipping beds is resumed 1 mile south of the Potomac River.

Southward to Leesburg, Virginia, the beds dip uniformly eastward, but the Triassic border fault gradually cuts out the entire Chilhowee group and the overlying Tomstown and Frederick limestones so that south of Leesburg the Catoctin volcanic rocks are thrown against the Triassic sediments.

Triassic Border Fault

The Triassic border fault along the eastern side of Catoctin Mountain, separates the Cambrian sequence of the South Mountain anticlinorium from the thick cover of Triassic sediments in the Triassic basin of the Frederick Valley. Poor exposures and a cover of alluvial wash from the slopes of Catoctin Mountain make mapping of the trace of the fault difficult. The writer's map of the Triassic border fault (Pl. 1) agrees for the most part with the map of Stose and Jonas (1938), but there are a few discrepancies.

Between Mount St. Marys and Catoctin Furnace in northern Maryland, Stose and Stose (1946, p. 89, Fig. 18) show the Triassic border fault consisting of a main fault between the Harpers phyllite and Frederick limestone and an additional branch fault between the Frederick limestone and the Triassic Gettysburg shale.

Since the stratigraphic relation between the Harpers phyllite and the Frederick limestone seems due either to an unconformity or a facies change, the writer sees no evidence for placing a fault here. Instead, the author maps (Pl. 1) only one fault between the Frederick limestone and the Gettysburg shale. The Frederick limestone is exposed in a belt parallel to U. S. Highway 40. This belt separates two large bodies of Triassic sediments, one extending northward into Pennsylvania, the other southward into Virginia. Both the writer (Pl. 1) and Stose and Jonas (1938) place the Triassic fault between the Frederick limestone and the Tomstown dolomite and thus continue the trend of the border fault parallel to Catoctin Mountain. An alternative interpretation may be that the Triassic fault is absent in this area so that the Tomstown and Frederick make a normal sequence which has not been downfaulted with the remainder of the Cambrian formations. This interpretation would imply that this portion of the Cambrian terrain stood above the surrounding downfaulted area and that Triassic sediments have lapped onto this higher area.

Petrofabrics

Introduction

A petrofabric analysis was undertaken to try to discover the relation of folds to grain orientation. Three oriented samples were taken from the upper limb, crest, and overturned limb of one anticline in the Weverton quartzite (Fig. 9, A, B, C). The fold shows no visible flow cleavage although a thickening of the beds near the crest and a thinning in the limbs is obvious. Six petrofabric diagrams (3 quartz and 3 mica) have been prepared. Thin sections were cut in the *ac* plane, and *B* is in the center of the diagrams.

FIGURE 8.—STRATIGRAPHIC RELATIONSHIPS OF WEVERTON

Showing lithologic units, variations in thickness within the Weverton, and stratigraphic relationship of Weverton to overlying Harpers phyllite and underlying Loudoun and Catoctin formations.

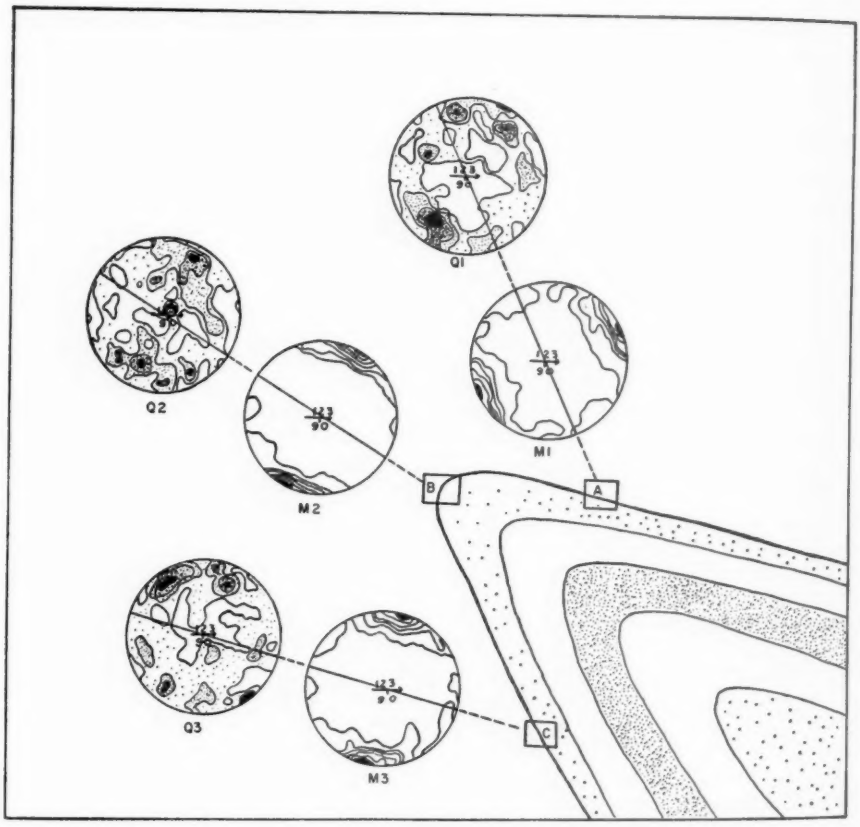

FIGURE 9.—PETROFABRIC DIAGRAMS
Showing their orientation in an anticline.

Specimen	*Diagram*	*Mineral*	*No. of grains*	*Contours (per cent)*
A	M1	mica	100	1, 3, 5, 7, 9, 11
B	M2	mica	100	1, 4, 6, 10, 14, 15
C	M3	mica	100	1, 4, 6, 10, 14, 15
A	Q1	quartz	200	1, 2, 3, 4
B	Q2	quartz	200	1, 2, 3, 4
C	Q3	quartz	200	1, 2, 3, 4

Mica Orientation

The three mica diagrams (M^1, M^2, M^3) show pronounced maxima in the position of the pole of the flow-cleavage plane. Hence, although no flow cleavage is visible in the field it is clear in the petrogabric diagrams. The different orientations of the maxima for positions *A*, *B*, and *C* in the fold clearly indicate that the flow cleavage fans gently westward. The existence of only one maxima in each diagram suggests that there is only one generation of *S* planes.

Quartz Orientation

According to the fracture hypothesis of Sander (1930), quartz grains are fractured and reduced to needlelike fragments which are rotated in the *ab* plane with their long axes parallel to *a*, the direction of movement. Sander assumed that the crystal axes of the quartz paralleled the long axes of the needles. Later experimental work by Griggs and Bell (1938) confirmed Sander's idea of the formation of needles by fracture but found that the

axes of the quartz needles paralleled not only the vertical edges of the crystal but also the rhombohedral and horizontal edges.

The position of the *ab* plane in diagrams Q^1, Q^2, and Q^3 (Fig. 9) is known from the mica diagrams. The maxima do not exceed 4 per cent and may, therefore, be unreliable for tectonic analysis. The 4 per cent maxima in Q^1 and Q^2, which are surrounded by exceptionally large 3 per cent maxima, correspond to the theoretical quartz maxima where fracturing takes place along the unit rhombohedron. The other 4 per cent maxima do not correspond to the theoretical maxima derived from fracturing parallel to either the vertical or horizontal edges of the quartz crystal.

Conclusions

The following conclusions are drawn from the present study:

(1) Catoctin Mountain is the eastern and upper limb of the South Mountain anticlinorium and is not a syncline as previously interpreted.

(2) What has previously been considered Loudoun on the eastern side of the mountain is a sequence including the upper Weverton and basal Harpers formations.

(3) Rhyolite tuffs in the Middletown Valley which were previously considered Loudoun are thought to be interbedded with the Catoctin metabasalt flows.

(4) No significant hiatus seems to exist between the metavolcanics and the overlying Chilhowee group; volcanic activity probably commenced with the Swift Run formation, reached its culmination as the Catoctin metabasalt flows and interbedded tuffs, and waned during Chilhowee time.

(5) Changes in facies within the Lower Cambrian clastics take place over a limited geographical area.

(6) The megascopic and microscopic analysis of folding indicates that shear folds result where the principal movement occurs along flow-cleavage rather than along bedding planes.

(7) Axial-plane cleavage dominates the area but parallels the axial plane only in the crests and troughs of folds.

(8) Flow cleavage forms a fan which opens toward the crests of anticlines and troughs of synclines; since the folds are asymmetrically overturned westward, the cleavage dips constantly to the east.

(9) Lineation in *a* results from a stretching of the minerals and a crenulation of the flow-cleavage planes.

References Cited

Amsden, T. W., 1951, Paleontology of Washington County: Md. Dept. Geology, Mines, Water Res., Physical features of Washington County, p. 98–123.

Barth, T. F. W., 1952, Theoretical petrology: N. Y., John Wiley and Sons, 386 pages.

Barrow, G., 1893, On an intrusion of muscovite-biotite gneiss in the southeastern Highlands of Scotland and its accompanying metamorphism: Geol. Soc. London Quart. Jour., v. 49, p. 330–358.

Bascom, F., 1896, The ancient volcanic rocks of South Mountain, Pennsylvania: U.S. Geol. Survey Bull. 136, p. 1–124.

Bassler, R. S., 1919, Report on the Cambrian and Ordovician formations of Maryland: Md. Geol. Survey, v. 19, 373 pages.

Bloomer, R. O., and Bloomer, R. R., 1947, The Catoctin formation of Central Virginia: Jour. Geology, v. 55, p. 94–106.

Cloos, E., 1941, Geologic map of Washington County: Md. Geol. Survey, 1:62,500.

—— 1946, Lineation, a critical review and annotated bibliography: Geol. Soc. America Mem. 18, 122 pages.

—— 1947, Oölite deformation in the South Mountain fold, Maryland: Geol. Soc. America Bull., v. 58, p. 843–918.

—— 1951, Stratigraphy of sedimentary rocks of Washington County: Md. Dept. Geology, Mines, Water Res., Physical features of Washington County, p. 17–94.

Eskola, P., 1920, The mineral facies in rocks: Norsk. Geol. Tidssk. 6, p. 142–194.

—— 1939, Die Entstehung der gestine: (Barth, T. F. W., Correns, G. W., and Eskola, P.), Springer, Berlin, 422 pages.

Geiger, H. R., and Keith, A., 1891, The structure of the Blue Ridge near Harpers Ferry: Geol. Soc. America Bull., v. 2, p. 155–164.

Goddard, E. N., *et al.*, 1951, Rock-color Chart: Geol. Soc. America.

Griggs, D., and Bell, J., 1938, Experiments bearing on the orientation of quartz in deformed rocks: Geol. Soc. America Bull., v. 49, p. 1723–1746.

Harker, A. C., 1939, Metamorphism: London, Methuen and Co. Ltd., 362 pages.

Howell, B. F., *et al.*, 1944, Correlation of the Cambrian formations of North America: Geol. Soc. America Bull., v. 55, p. 993–1003.

Jonas, A. I., and Stose, G. W., 1939, Age relations of the Precambrian rocks in the Catoctin Mountain-Blue Ridge and Mount Rogers anticlinoria in Virginia: Am. Jour. Sci., v. 237, p. 575–593.

Keith, A., 1892, Geologic Structure of the Blue Ridge in Maryland and Virginia: Am. Geologist, v. 10, p. 362–368.

—— 1893, Geology of the Catoctin belt: U.S. Geol. Survey, 14 Ann. Rept., pt. 2, p. 285–395
— 1894, Harpers Ferry Folio: U.S. Geol. Survey. Atlas no. 10.
Keyes, C. R., 1890, Discovery of fossils in limestone of Frederick County, Maryland: Johns Hopkins Univ. Circ., v. 10, p. 32.
King, P. B., 1949, The base of the Cambrian in the southern Appalachians: Am. Jour. Sci., v. 247, p. 512–530, 622–645.
Nevin, C. M., 1949, Principles of structural geology: N. Y., John Wiley, 4th ed., 410 pages.
Resser, C. E., 1938, Cambrian system (restricted) of the Southern Appalachians: Geol. Soc. America Spec. Papers, no. 15, 140 pages.
Safford, J. M., 1856, A geological reconnaissance of the state of Tennessee: Tenn. Geol. Survey 1st Bienn. Rep. 164 pages.
Sander, B., 1930, Gefugekunde der Gesteine: Julius Springer, Wein, 352 pages.
Scotford, D. M., 1951, Structure of the Sugarloaf Mountain area as a key to Piedmont stratigraphy: Geol. Soc. America Bull., v. 62, p. 45–76.
Snyder, F. G., 1947, The problem of the Lipalian interval: Jour. Geology, v. 55, p. 146–152.
Stose, G. W., 1906, The sedimentary rocks of South Mountain, Pennsylvania: Jour. Geology, v. 14, p. 201–220.
Stose, G. W., and Jonas, A. I., 1935, Limestone of the Frederick Valley, Maryland: Wash. Acad. Sci. Jour., v. 25, p. 564–565.
Stose, G. W., and Stose, A. J., 1946, Geology of Carroll and Frederick counties: Md. Dept. Geology, Mines, Water Res., p. 11–131.
Turner, F. J., 1948, Mineralogical and structural evolution of metamorphic rocks: Geol. Soc. America Mem. 30, 342 pages.
Ulrich, E. O., and Cooper, G. A., 1938, Ozarkian and Canadian brachiopods: Geol. Soc. America Spec. Paper no. 13, 323 pages.
Ulrich, E. O., and Foerste, A. F., 1935, New genera of Ozarkian and Canadian Cephalopods: Denison Univ. Bull., Jour. Sci. Lab., v. 30, p. 259–303.
Walcott, C. D., 1896, The Cambrian rocks of Pennsylvania: U. S. Geol. Survey Bull. 134, 43 pages.
Wheeler, H. F., 1947, The base of the Cambrian system: Jour. Geology, v. 55, p. 153–159.
Williams, G. H., and Clark, W. B., 1893, Maryland, its resources, industries, and institutions: Baltimore, Sun Job Printing Office, 504 pages.
Wilson, J. L., 1952, Upper Cambrian stratigraphy in the central Appalachians: Geol. Soc. America Bull., v. 63, p. 275–322.

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Short Notes

PILLAR FALLS MUD FLOW AND SHOSHONE FALLS ANDESITE NEAR TWIN FALLS, IDAHO

BY HAROLD T. STEARNS

The writer mapped and named the Pillar Falls mud flow and the underlying Shoshone Falls andesite in Snake River Canyon near Twin Falls, Idaho (Stearns *et al.*, 1938, p. 37, 38). The andesite is more than 200 feet thick, and a recent re-examination of the Pillar Falls mud flow indicates that it is the typical block lava top of the andesite rather than a mud flow. The fine-grained material filling the interstices of the blocky lava phase is clay derived in part from the decomposition of the original finely fragmented glassy andesite and in part from infiltered material from the sediments subsequently deposited on the rough loose surface. The blocky crustal phase of the flow is commonly 50 feet thick and is mostly dense glassy porphyritic andesite. The usual pumiceous crustal blocks are absent. As much as 2 feet of sandy sedimentary debris containing water-worn red vesicular basalt pebbles lies on the eroded surface of the blocky phase in the south bank of Snake River at Shoshone Falls. The overlying Pleistocene basalts rest on a smooth erosion surface truncating the blocky lava, indicating that the blocky lava was firmly cemented at the time of erosion.

The andesite was considered Miocene (?) because of its similarity to Tertiary silicic lavas in the Mount Bennett area (Piper, 1926, p. 8). A recent examination of the lavas in the Mount Bennett area indicates that silicic lavas there are deeply weathered whereas many of the small glass fragments of the crustal phase of the Shoshone Falls andesite are only slightly weathered, suggesting Pleistocene age. However, glassy silicic lavas are remarkably well preserved elsewhere in the Miocene (?) rocks as shown by a silicic lava flow with its original crescentic surface flow pattern now partly exhumed by erosion of the overlying Pliocene lake beds near the head of Castle Creek southeast of Murphy, Idaho.

The Pliocene Banbury basalts only 3 miles downstream from the Shoshone Falls andesite are deeply weathered. In fact, they are more deeply weathered than most of the Columbia River basalts of Miocene age, a condition possibly accelerated by the moist environment in ancient Lake Idaho. Lava flows interbedded with the Idaho sediments and filling deep valleys cut during the deposition of the lake beds have been found in both the Bliss and the Bruneau areas in Idaho, attesting that the Idaho formation was not deposited by a single lake; yet nowhere can the typical Columbia River lavas be found interbedded with the Idaho formation. Therefore, very different rates of weathering must have occurred in this part of Idaho during the late Cenozoic.

REFERENCES CITED

Piper, A. M., 1926, Ground water for irrigation on Camas Prairie, Camas and Elmore Counties, Idaho: Idaho Bur. Mines and Geology, Pamphlet 15.

Stearns, H. T., Crandall, Lynn, and Stewart, W. G., 1938, Geology and ground-water resources of the Snake River Plain, Idaho; U. S. Geol. Survey W.-S. Paper 774.

HOPE, IDAHO

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TORO SEAMOUNT

BY J. LAMAR WORZEL AND G. LYNN SHURBET

On the basis of submarine topographic echoes, Luskin *et al.* (1952) predicted a seamount (designated G in that work) at about 33° 22′ N., 60° 48′ W. This was the only one of 36 echoes which could not be accounted for by known topography. The echoes result from the reflection of explosion-generated sounds by seamounts, submarine ridges, or continental slopes which block or partially block the natural sound channel in the ocean (Ewing and Worzel, 1948). Many of these sounds have travelled more than 1000 miles.

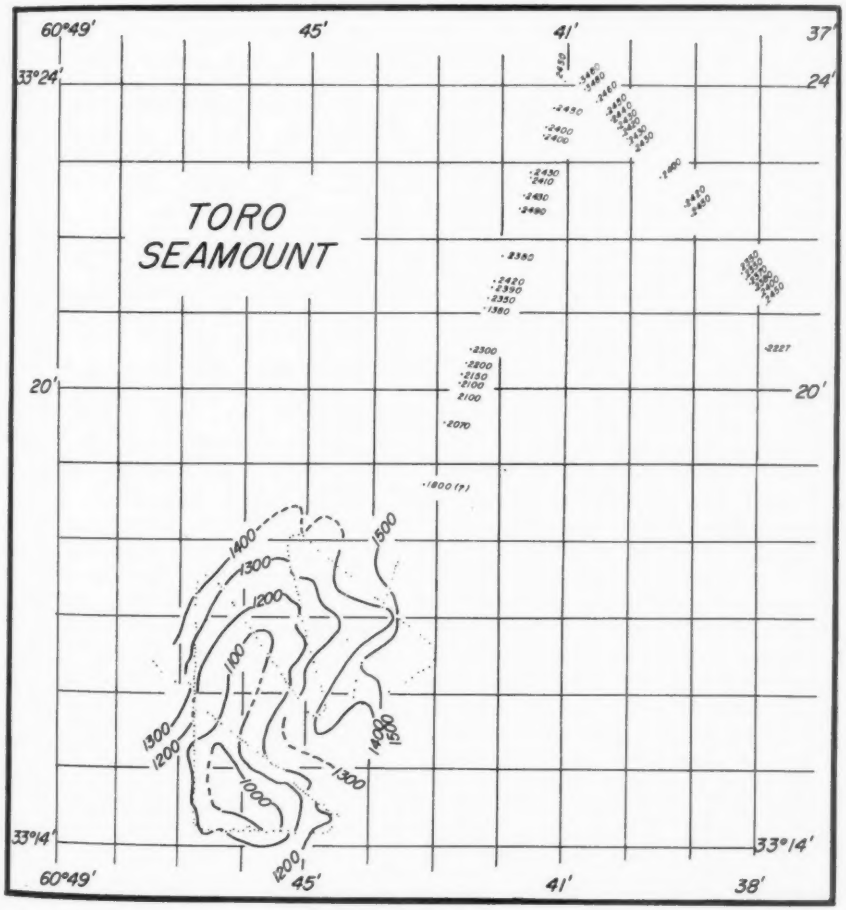

FIGURE 1.—TORO SEAMOUNT

In March 1952 USS TORO (SS 422) was operating near the predicted seamount. In the 5 hours available for sounding the area, a seamount was found 7 miles southeast of the

predicted location. We propose to name it Toro Seamount following the tradition of naming seamounts for vessels.

Eight lines of soundings (indicated by dots in Fig. 1) were made to determine the minimum depth and the shape of this seamount. The submarine surfaced after dark, so evening celestial observations were not available prior to the start of the search. Loran reception was too poor to obtain a fix until about 2 hours after departing from the area, so the absolute position is believed to be reliable only to about 3 miles. A Dead Reckoning Tracer was used in the navigation, and the relative positions over the seamount are probably accurate to within about half a mile.

It is interesting to note that the predicted position is in disagreement with the observed position by only 1° in bearing and 3 miles in distance from Bermuda.

The soundings were made visually with the standard NGA-1 echo-sounding equipment adjusted to a sound velocity of 4800 feet per second. No corrections to the soundings have been made for the velocity variations of the water column or for the effect of slope.

The minimum sounding observed was 980 fathoms on the southeast side of the sounding area. The peak was not completely delineated by the soundings, but an L-shaped peak was indicated. This does not suggest a single volcanic cone, but could be the merging of two or more such cones.

While clearance to publish this note was being obtained, Northrop and Frosch (1954) obtained clearance and published their sounding data, taken several months later, over what is probably a part of the same seamount reported here. Their paper reported Seamount George (as called in that paper) at 33° 19′N., 60° 49′ W., with a minimum sounding of 1210 fathoms. In the present paper, Toro Seamount is reported at 33° 15′ N., 60° 46′ W., 5 miles southeast of their Seamount George, with a minimum sounding of 980 fathoms. The difference in reported positions is within the limits of accuracy claimed for the positions; however the minimum sounding of 1210 fathoms for Seamount George is not compatible with soundings reported in the present paper for Toro Seamount where a large area of soundings less than 1200 fathoms was determined. It is probable that the peak we report is southeast of the double peak which they report, in the area where they have poor sounding control. We conclude this must be a compound feature which combines at least three peaks. It must present a broad face (at least 10 miles in length with possible extensions to the southeast) toward Bermuda, which would account for the strong topographic echoes recorded at Bermuda by Luskin *et al.* (1952).

We wish to express our sincere appreciation to Captain Earl T. Hydeman, Commander Submarine Development Group Two, and Commander Gordon K. Nicodemus, Jr., Commanding Officer, USS Toro, who made this work possible. All the officers and men of USS Toro co-operated fully to make this project successful. This work was carried out under contract N6 onr 271 TO 8 and TO 24 with the Office of Naval Research, Department of the Navy.

References Cited

Ewing, Maurice, and Worzel, J. Lamar, 1948, Long-range sound transmission *in* Ewing, Maurice, Worzel, J. Lamar, and Pekeris, C. L., Propagation of sound in water: Geol. Soc. America Mem. 27, p. 1–35.

Luskin, Bernard, Landisman, Mark, Tirey, G. B., and Hamilton, G. R., 1952, Subtopographic echoes from explosives sound: Geol. Soc. America Bull., v. 63, p. 1053–1068.

Northrop, John, and Frosch, Robert A., 1954, Seamounts in the North American Basin: Deep-Sea Research, v. 1, no. 4, p. 252–257.

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DISCOVERY OF *SONORASPIS* IN SOUTHERN CALIFORNIA

BY ALEXANDER STOYANOW AND TAKEO SUSUKI

Genus *Sonoraspis* was erected by the senior writer (Stoyanow, 1952, p. 50, Pl. 14, figs. 1-6) for the Middle Cambrian trilobites from Sonora, Mexico, which differ from *Anoria* Walcott (1924, p. 54; 1925, p. 67) and *Glossopleura* Poulsen (1927, p. 268), under which generic names they had been originally listed (Stoyanow, 1942, p. 1263–1264), in having eight instead of seven thoracic segments and in possessing mixed characteristics of those two genera.

The discovery of Cambrian trilobites in Sonora led McKee (1947, p. 282) to postulate an early Middle Cambrian seaway from the Grand Canyon area to Sonora across southwestern Arizona. This view was based chiefly on the presence of *Glossopleura* in Harquahala Mountains south of the Grand Canyon. The above interpretation was opposed by the senior writer (Stoyanow, 1948, p. 323–324) who pointed out the absence of Cambrian strata between the two regions, and especially the fact that in Cambrian time southwestern Arizona was occupied by the Precambrian massive, Altar Headland (Stoyanow, 1942, p. 1264) of Mazatzal Land (Stoyanow, 1936, p. 462; *cf.* Cooper and Arellano, 1946, p. 608), which probably extended southward into Sonora. A westward Paleozoic portal in the Sonora basin was suggested.

A trilobite of the *Anoria-Glossopleura* group but with eight thoracic segments, collected from the Middle Cambrian strata in Bristol Mountains, southern California, was described as *Bathyuriscus howelli* var. *lodensis* by Clark (1921, p. 6) but was not figured. Later Resser (1928, p. 10, Pl. 3, fig. 9) illustrated Clark's type advancing it to the status of species, *Dolichometopus? lodensis* (Clark), and also mentioned eight thoracic segments. However, in a more recent paper Resser (1935, p. 29, 34) placed this species, as *Glossopleura lodensis* (Clark), with other species possessing only seven thoracic segments, as also did Mason (1935, p. 109, Pl. 15, fig. 11) while describing, as *Anoria lodensis*, a plesiotype of Clark's species collected in Marble Mountains. Like *Anoria*, Clark's species has a macropleural fifth thoracic segment. The illustrations of Resser and Mason do not allow the figuring of the number of thoracic segments (eight as by Clark, or seven as by Resser and Mason) in either of two types with any degree of assurance. Clark's holotype was not available at the time of this writing, but Mason's plesiotype, preserved in the Los Angeles Museum, has been examined by the writers through the kindness of Dr. G. P. Kanakoff. It has seven thoracic segments with the fifth segment macropleural as in *Anoria*. Since the thoracic axis is completely destroyed in this specimen, it is impossible to establish the presence or absence of the axial tuberculation.

The possible presence of *Sonoraspis* in southern California, as suggested by Clark's mention of eight thoracic segments, stimulated a more detailed research in the Middle Cambrian strata of California. Through indefatigable and systematic field studies the junior writer eventually located *Sonoraspis* in Marble Mountains about 50 feet below the base of Bonanza King formation. Like *S. torresi* and *S. gomezi* of Sonora, it has, besides the presence of eight thoracic segments of the generic diagnosis, and like those two species, certain mixed characteristics of *Anoria* and *Glossopleura*. *Sonoraspis californica*, sp. nov., has tubercles on the anterior thoracic segments and the course of the facial sutures as in *Anoria*, but like *Glossopleura* it lacks the macropleural enlargement of the fifth thoracic segment. If Clark's species from Bristol Mountains has eight segments in the thorax, it is distinguished by the presence of the latter characteristic. The diagnostic description of the new species is presented below.

Sonoraspis californica, sp. nov.

(Plate 1, figures 1 and 2)

This species has eight thoracic segments (none macropleural) as in the genotype, *S. torresi* (untuberculated thoracic segments and facial sutures as in *Anoria*), and as in *S. gomezi* (tuberculated thoracic segments and facial sutures as in *Glossopleura*) from Sonora. It differs from the first species in having tubercles on the thoracic segments and from the second by the *Anoria*-like course of the facial sutures which is as in *S. torresi*. It is further distinct from *S. gomezi* by the presence of tubercles in the anterior segments of the thorax.

In holotype the tubercles of two posterior segments are destroyed, but the paratype, although with a damaged median line, clearly shows that all thoracic segments were tuberculated, as in *Anoria*.

Diagnostic differences: from *Anoria* and *Glossopleura*, by eight thoracic segments; from *S. torresi*, by the thoracic tuberculation; from *S. gomezi*, by the course of facial sutures.

Type: Holotype: U.C.L.A. Pal. Coll., Cat. No. 23460; Paratype: U.C.L.A. Pal. Coll., Cat. No. 23461.

Occurrence: In bedded buff to yellow limestone, Cadiz formation, 50 feet below the base of Bonanza King formation, Marble Mountains, San Bernardino County.

The paleogeographic significance of the described find is as follows. The presence of either *Anoria* or *Glossopleura* in Sonora has not been established. The four figured specimens from the Cambrian of Sonora with the thorax of eight segments preserved, other than those described by the senior writer, which were discussed under *Glossopleura* by Lochman (1952, p. 137, Pl. 31, figs. 8, 9, 12, 14), all belong in *Sonoraspis*. There is not a single thorax illustrated or mentioned in the text within the assemblage of cranidia and pygidia described by Lochman (1952, p. 135, Pl. 25, figs. 1–21) as *Glossopleura leona*. All these specimens may as well belong in *Sonoraspis*. The logical conclusion might be that the *Sonoraspis* fauna was endemic to the Sonora basin, but the find in Marble Mountains described here clearly indicates intermigration with southern California. No eight-segmented forms of the *Anoria-Glossopleura* group have been reported from the Cambrian of the Grand Canyon thus far, but along the main Cordilleran trough such specimens appear to be present as seen from Walcott's illustrated material from Canada and (less certainly) from Utah (Stoyanow, 1952, p. 50–51).

The Marble Mountains, where the *Sonoraspis* beds have been located, are in San Bernardino County between Barstow and Needles, about 2 miles northeast of Chambless on U. S. Highway 66, 90 miles east of Barstow. This locality has been described twice. In an earlier paper, Hazzard and Crickmay (1933, p. 60, 64, 65, 68) placed in the Middle Cambrian about 370 feet of strata between the topmost Lower Cambrian layer with Mesonacidae and the gray massive and bedded limestone complex above which was tentatively interpreted as Goodspring formation. The index fossil collected in the upper 105 feet of the Middle Cambrian, thus defined, was identified as *Dolichometopus productus* (H. and W.). In the second paper (Hazzard and Mason, 1936, p. 232, 233) the combined Middle Cambrian sequence of Providence and Marble Mountains was termed Cadiz formation, whereas the tentative Goodspring formation was changed to Bonanza King formation and added to the Middle Cambrian as also tentatively was the overlying Cornfield Spring formation. The fauna from the upper 105 feet of Cadiz formation was identified and listed. A collection from Marble Mountains was described by Mason (1935). The Cadiz fauna, as

PLATE 1.—*SONORASPIS CALIFORNICA*, N.SP. AND COLLECTING GROUNDS

FIGURE 1.—HOLOTYPE. U.C.L.A. PAL. COLL., CAT. NO. 23460. × 1.8

FIGURE 2.—PARATYPE. U.C.L.A. PAL. COLL., CAT. NO. 23461. × 2.1

FIGURE 3.—GENERAL VIEW OF MARBLE MOUNTAINS FROM U. S. HIGHWAY 66

FIGURE 4.—WESTERN PART OF MARBLE MOUNTAINS

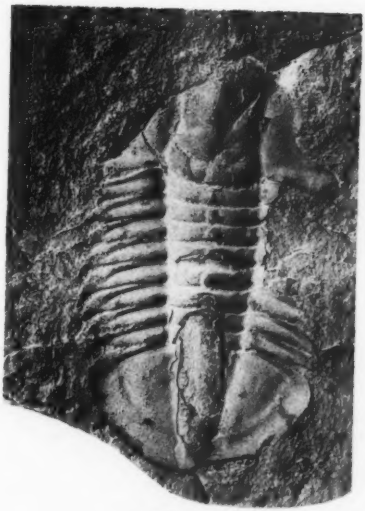

FIGURE 1

FIGURE 2

FIGURE 3

FIGURE 4

SONORASPIS CALIFORNICA, N. SP., AND COLLECTING GROUNDS

FIGURE 1

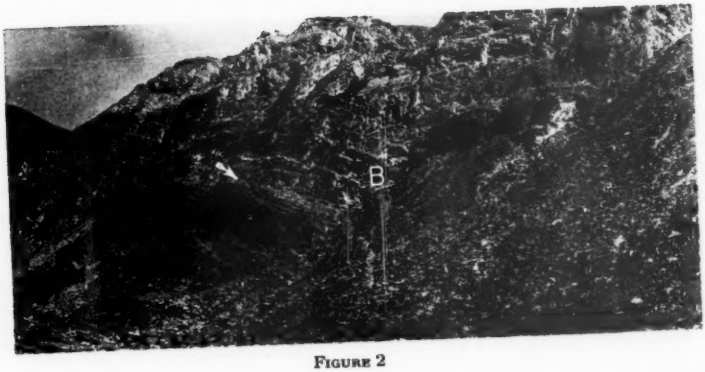

FIGURE 2

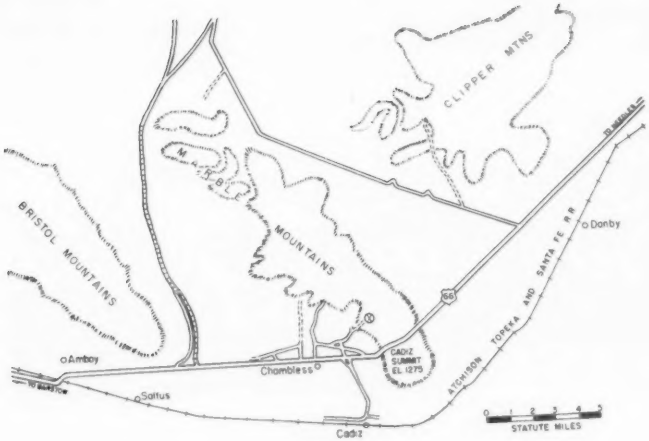

FIGURE 3

VIEWS OF COLLECTING LOCALITY

listed, apparently represents several individual horizons.

It is safe to conclude that the strata with *Sonoraspis* in Marble Mountains are correlatively within 500 feet of Arrojos formation in Sonora, which is the vertical range of *Sonoraspis* in that formation, and within which this trilobite genus occurs between the zones of *Arellanella* and *Zacanthoides* (*cf.* Cooper and Arellano, 1952, p. 6; location of Arrojos Hills is indicated as "800" south of the southern bend of the Rio Magdalena, see map on page 3 of the same publication).

Addendum

The recent paper of A. R. Palmer (1954) on the appraisal of Middle Cambrian trilobites from Great Basin was received by the writers after the present article had been submitted for publication. Palmer's text concerning *Sonoraspis* reads (p. 67):

"*Sonoraspis* (Stoyanow, 1952) is based on poorly preserved material that differs from *Glossopleura* by having 8 instead of 7 thoracic segments. The writer does not consider this sufficient justification for generic distinction, especially with inadequate knowledge available concerning variation of this feature within a population. *Sonoraspis* is considered here to be a synonym of *Glossopleura*.

The present writers are of opinion that: (1) the qualitative characteristics of Cambrian trilobites by far outweigh in significance the qualitative features (so profusely used for generic and specific diagnoses in the post-Walcottian literature); (2) the genus *Sonoraspis* was erected not only on the additional thoracic segment, but also on the comingling of characteristics separately belonging to *Anoria* and *Glossopleura*, whereas no such combinations have been demonstrated for the above genera; (3) the principal task of any paleontologist is in the application of his ability to derive correct interpretation from impaired materials, and deductions made by the senior writer from the less perfect Sonora material have been corroborated by the excellent specimens from California; (4) there is a considerable range of intrageneric variations (including spinosity) in *Sonoraspis*, but the eighth thoracic segment invariably is present.

References Cited

Clark, C. W., 1921, Lower and Middle Cambrian formations of the Mohave Desert: Univ. Calif., Dept. Geol. Sci. Bull., v. 13, p. 1–7.

Cooper, G. A., and Arellano, A. R. V., 1946, Stratigraphy near Caborca, northwest Sonora, Mexico: Am. Assoc. Petroleum Geologists Bull., v. 30, p. 606–611.

—— 1952, Introduction and stratigraphy *in* Cambrian stratigraphy and paleontology near Caborca, northwestern Sonora, Mexico: Smithson. Misc. Coll., v. 119, no. 1, p. 1–23, Pls. 1–5.

Hazzard, John C., and Crickmay, Colin H., 1933, Notes on the Cambrian rocks of the eastern Mohave Desert, California: Univ. Calif., Dept. Geol., Bull., v. 23, no. 2, p. 57–78, Pl. 1.

Hazzard, John C., and Mason, John F., 1936, Middle Cambrian formations of the Providence and Marble Mountains, California: Geol. Soc. America Bull., v. 47, p. 229–240.

Lochman, Christina, 1952, Trilobites *in* Cambrian stratigraphy and paleontology near Caborca, northwestern Sonora, Mexico: Smithson. Misc. Coll., v. 119, no. 1, p. 60–161, Pls. 18–31.

Mason, John F., 1935, Fauna of the Cambrian Cadiz formation, Marble Mountains, California: Southern Calif. Acad. Sci. Bull., v. 34, pt. 2, p. 97–119, Pl. 15.

McKee, E. D., 1947, Paleozoic seaways in western Arizona: Am. Assoc. Petroleum Geologists Bull., v. 31, p. 282–292.

Palmer, Allison R., 1954, An appraisal of the Great Basin Middle Cambrian trilobites described before 1900: U. S. Geol. Survey Prof. Paper 264-D, p. 55–86, Pls. 13–17.

Poulsen, Chr., 1927, The Cambrian, Ozarkian and Canadian faunas of northwest Greenland: Meddelelser om Grønland, v. 70, p. 233–343, Pls. 14–21.

Resser, Charles E., 1928, Cambrian fossils from the Mohave Desert: Smithson. Misc. Coll., v. 81, no. 2, p. 1–14, Pls. 1–3.

——, 1935, Nomenclature of some Cambrian trilo-

Plate 2.—VIEWS OF COLLECTING LOCALITY

Figure 1.—Collecting Grounds, Marble Mountains, San Bernardino County, California

T—indicates the trail

Figure 2.—Close View of Collecting Locality

Arrow indicates *Sonoraspis* beds. B—marks the basal part of Bonanza King formation.

Figure 3.—Orientation Map

bites: Smithson. Misc. Coll., v. 93, no. 5, p. 1–46.

Stoyanow, Alexander, 1936, Correlation of Arizona Paleozoic formations: Geol. Soc. America Bull., v. 47, p. 459–540, Pl. 1.

—— 1942, Paleozoic paleogeography of Arizona: Geol. Soc. America Bull., v. 53, p. 1255–1282, Pls. 1–5.

—— 1948, Some problems of Mississippian stratigraphy in southwestern United States: Jour. Geology, v. 56, p. 313–326.

—— 1952, The original collection of Cambrian trilobites from Sonora *in* Cambrian stratigraphy and paleontology near Caborca, northwestern Sonora, Mexico: Smithson. Misc. Coll., v. 119, no. 1, p. 49–59, Pl. 14.

Walcott, Charles D., 1924, Cambrian geology and paleontology. Cambrian and Ozarkian trilobites: Smithson. Misc. Coll., v. 75, no. 2, p. 53–60, Pls. 9–14.

—— 1925, Cambrian geology and paleontology. Cambrian and Ozarkian trilobites: Smithson. Misc. Coll., v. 75, no. 3, p. 61–146, pls. 15–24.

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